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Supracrustal Sequences of the Belomorian Mobile Belt: Protoliths, Age, and Origin

T. A. Myskova, V. A. Glebovitskii, Yu. V. Miller, A. B. L'vov†, A. B. Kotov, V. P. Kovach, and N. Yu. Zagornaya

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Abstract—Two section types are established in the Chupa nappe of the Early Precambrian Belomorian mobile belt (BMB) of the Baltic shield: one in the Lake Verkhnee Pulongskoe area of the stratotype region (Chupa Complex) and the other one in the tectonic window of the Kovdozero nappe (Kalikorva Complex). Both complexes are composed of supracrustal rocks represented by equal proportions of metavolcanics and metasediments (amphibolites and gneisses). Metavolcanics constitute 5–10% of the Chupa Complex and more than 50% of the Kalikorva Complex. Tholeiitic basalts prevailing among metavolcanics are accompanied by dacites and dacitic andesites. Poorly sorted graywackes with a low chemical index of aging (CIA = 51–55) are protoliths of metasediments. The dating results obtained for detrital and metamorphic zircons from dacites and graywackes (Bibikova *et al.*, 2001), and the Sm–Nd isotopic-geochemical study of rocks imply coeval ages of the complexes and synchronism of volcanism and sedimentation. The supracrustal rocks were formed 2870 ± 30 Ma ago according to isotopic dates established for acidic volcanics (Bibikova *et al.*, 2001). The detrital material of graywacke sequences could be derived from both the constituent volcanics of the complexes and the proto-ophiolites of mafic zones. An insignificant admixture of older clastic material derived from a remote provenance (presumably from the Karelian craton) cannot be ruled out as well. Local differences in volcanism and sedimentation were probably responsible for compositional difference between the Kalikorva and Chupa complexes and their constituting paragneisses. A low maturity and substantial proportion of mafic material were characteristic of protoliths for metasediments. The supracrustal rocks likely formed in the immature island-arc setting.

Key words: geochemistry, Sm–Nd isotopic systematics, Chupa Complex, Kalikorva Complex, Late Archean, Baltic shield.

INTRODUCTION

According to recent data, supracrustal rocks of the Belomorian mobile belt (BMB) constituting not more than 20% of the total belt volume are mainly localized in the Chupa tectonic nappe (Glebovitskii *et al.*, 1996; Miller, 1997; Miller *et al.*, 1999). The nappe extends as a continuous series of separate structural exposures throughout the entire mobile belt from the Kem' area in the southeast to Kovdor in the northwest and farther to the state boundary between Russia and Finland (Fig. 1). The Chupa nappe rests, with a distinct structural unconformity, on the Khetolambina nappe composed predominantly of tonalites 2800 to 2740 Ma old, which enclose zones of mafic and ultramafic rocks 2880 to 2860 Ma old (Bibikova *et al.*, 1999) in some areas, while metagabbroids and, less commonly, metabasalts with rare paragneiss horizons are characteristic of the others. In opinion of many researchers (Stepanov and Slabunov, 1989; Miller and Mil'kevich, 1995; Miller, 1997; Lobach-Zhuchenko *et al.*, 1998; Bibikova *et al.*, 1999), the Khetolambina nappe includes tectonic slices of variably metamorphosed rocks of the Late Archean oceanic crust. The aim of this paper is to determine, using structural-geological, geochemical, and isotopic-geochemical criteria, the age limits for Chupa supracrustal sequences and to find their place in the Archean chronostratigraphic scale. Stratigraphic interpretation of the Chupa gneiss complex is important for understanding the dynamics of interaction between the BMB and Karelian craton.

GEOLOGICAL POSITION

In vast areas, the supracrustal rocks of the Chupa Complex are thrust, along with the underlying Khetolambina nappe, under the margin of the Karelian craton (a result of the Late Archean subduction), being exposed now in many tectonic windows (Miller and Mil'kevich, 1995). Judging from spatial distribution of tectonic windows, the underthrusting amplitude was as great as 80 km at least. The subduction time from 2850 to 2820 Ma is inferred from the age of the Tikshozero

† Deceased.

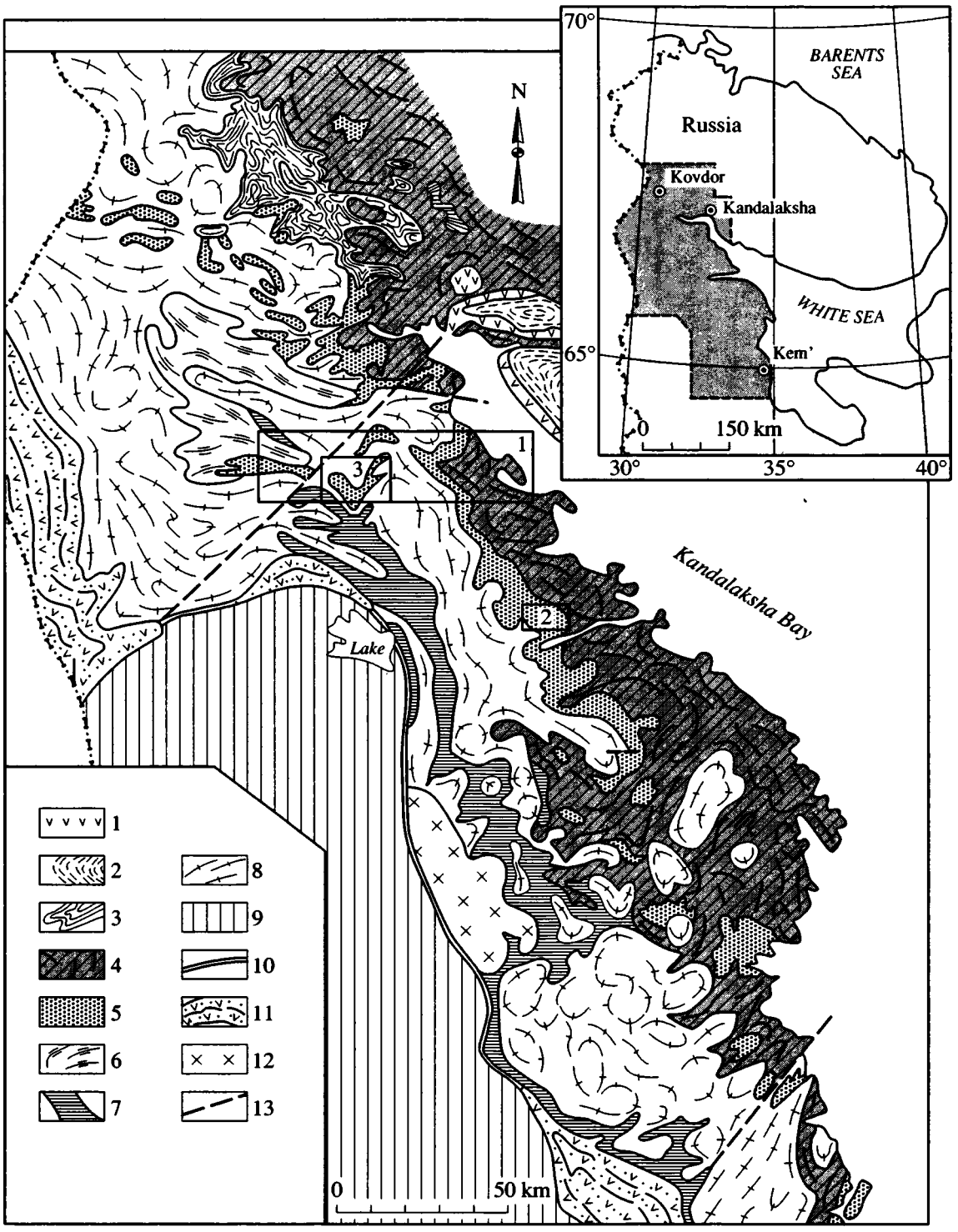

Fig. 1. Schematic tectonic map of the Belomorian mobile belt: (1) relicts of the Sumian rifting structures composed of metamorphosed basaltic andesites; (2) Laplandian (basic to intermediate granulites) and (3) Rikolatvino (tonalite gneisses with concordant bodies of mafic, ultramafic, and supracrustal rocks) nappes of the Svecofennian allochthon; (4) Khetolambina nappe (metamorphosed tholeiitic basalts and gabbroids injected by tonalites and intercalated with large concordant bodies of metabasites and ultramafics of mafic zones), (5) Chupa nappe (metamorphosed graywackes and subordinate intermediate, acidic, and less abundant basic metavolcanics), and (6) Orijarvi nappe (tonalite gneisses) of the Late Archean (Belomorian) allochthon; (7) Tikshozero (Keret') greenstone belt or the upper structural stage (tholeiitic metabasalts, intermediate and acidic metavolcanics) and (8) lower structural stage (tonalite-trondhjemite association) of the Kovdozero nappe; (9) Karelian craton, undivided; (10) boundary between the Belomorian mobile belt and Karelian craton and (11) troughs along this boundary with Lopian, Sumian, Jatulian, and Ludicovian rocks of the Svecofennian allochthon; (12) Topozero Complex of charnockites; (13) faults. Outlined rectangle areas are (1) Kovdozero geodynamic area, (2) Pulongskoe area (Fig. 2) and (3) Kalikorva area (Fig. 3).

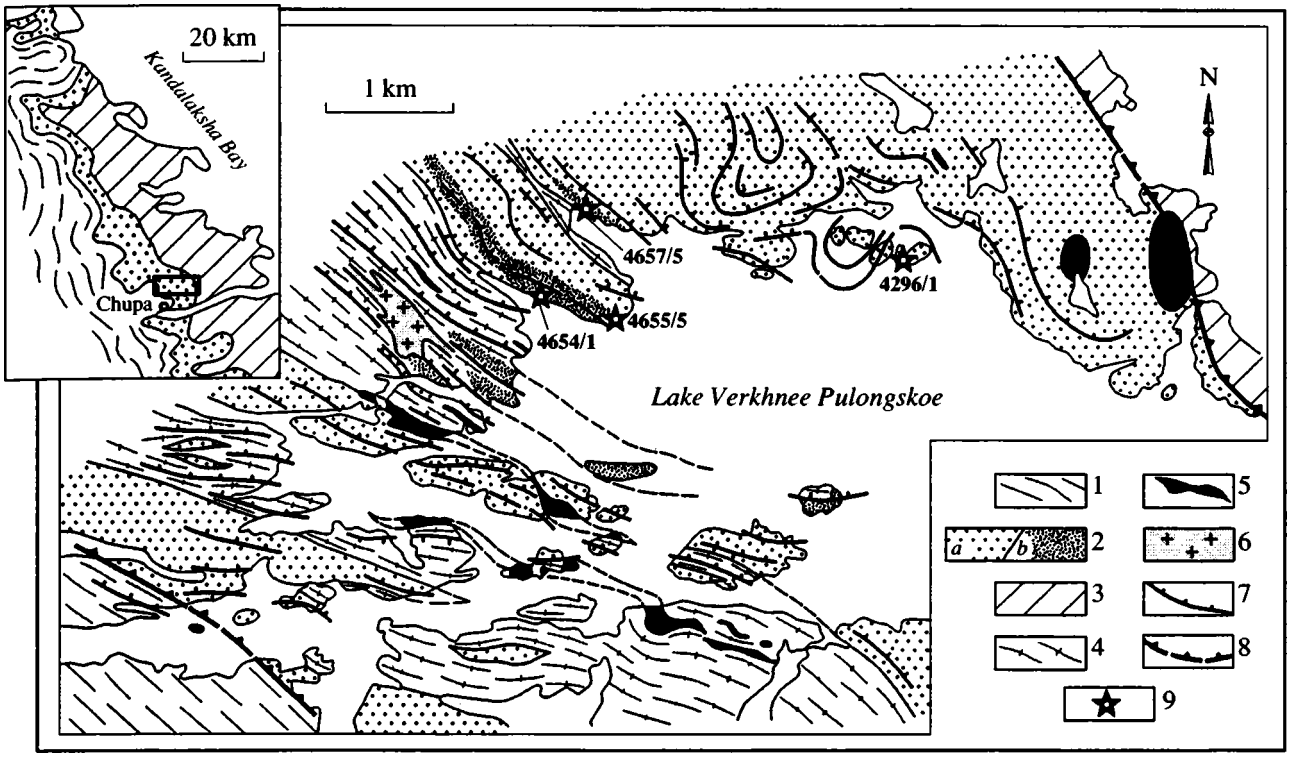

Fig. 2. Schematic geological structure of the Pulongskoe area: (1) Kovdozero nappe (tonalites); (2) Chupa nappe of variably migmatized coarse-grained kyanite-garnet-biotite gneisses (a) and similar rocks with abundant relicts of fine-grained garnet-biotite gneisses (b); (3) Khetolambina nappe (tonalites); (4) tonalites of the Chupa nappe; (5) Early Proterozoic gabbro-norites; (6) microcline granites; (7) generalized attitude of schistosity; (8) thrusts planes between nappes; (9) sampling sites for Sm–Nd isotopic-geochemical analysis and their numbers (inset map shows geographic position of the Pulongskoe area).

greenstone belt that marks the craton margin (Bibikova *et al.*, 1999). According to Mil'kevich and Miller (1999), geochemical characteristics of calc-alkaline volcanics from that belt suggest generation of their parental magmas during the partial melting of subducted oceanic crust under high temperatures and pressures.

Thus, the relatively thin (1–3 km) Chupa nappe composed of supracrustal rocks unconformably rests upon proto-oceanic crust (Khetolambina thrust) and is unconformably overlain by the margin of the Karelian craton, which is partly in the allochthonous position and represents, in a form of the Kovdozero thrust, a constituent of the BMB fold-and-thrust structure. Because of superimposed processes related to the Late Archean collision, Paleoproterozoic (Seletskian) rifting, and Svecofennian collision, which took place 2740–2690, 2350–2450, and 1950–1800 Ma ago, respectively, these relationships established in the Kovdor geodynamic area are substantially modified in other segments of the BMB. For instance, the entire packet of tectonic slices in the Chupa segment is overturned forming the giant recumbent isoclinal folds there. In other areas, the described relationships are substantially distorted by the Seletskian linear folds, Svecofennian thrusts and dome structures, and by other superimposed deformations.

Two areas of the BMB were selected for detailed studies of supracrustal rocks. One is the Pulongskoe area traditionally considered as a stratotype of the Chupa gneisses. In the other Kalikorva area, supracrustal rocks of the Chupa nappe are exposed in a synonymous tectonic window in the marginal part of the Karelian craton (2 and 3, respectively, in Fig. 1).

The Pulongskoe Area

In this area, the Chupa nappe studied along the profile approximately 8 km long is of a uniform composition and relatively simple, at first glance, structure, because schistosity, which coincides with metamorphic banding and locally with relicts of metamorphic lineation, dips southeastward almost everywhere and only sometimes is complicated by intricate folds (Fig. 2). Nevertheless, this is a seeming simplicity. The area is located between the Karelian (on the northwest) and Chupa (on the southeast) Svecofennian domes with systems of abundant mica-bearing pegmatites confined to their arches. In the study area, the thrust section is tripled, and its thickness estimated with due account for the average dip angle of rocks is approximately 2 km.

Intricate folds complicate the depressed dome structure, and non-migmatized rocks with relict bedding mostly occur southwest of the dome arch, where supracrustal rocks are overturned. In addition, Svecofennian domes are superimposed on the overturned limb of a large isocline, which determines the overturned position of the Chupa, Khetolambina, and Kovdozero nappes.

The Kalikorva Area

The area corresponds to the synonymous intricate antiform exposed in a tectonic window. Limbs of the antiform are composed of tonalite-gneisses of the Kovdozero nappe (the allochthonous marginal part of the Karelian craton), and supracrustal Chupa sequence is exposed in the antiform core. Deciphering folded structure of the Kalikorva antiform, we reconstructed the section of supracrustal rocks, which appeared to be compositionally more diverse and variable than the Pulongskoe area section (Fig. 3). The basal Member 1 of the section is composed of garnet amphibolites (up to 95%) with rare intercalations of biotite and biotite-amphibole gneisses. The member encloses lenses of garnet-biotite gneisses up to 20 m thick. The apparent thickness of Member 1 is over 600 m. Member 2 consists of garnet-biotite and less common biotite gneisses with subordinate intercalations (10%) of amphibolites and amphibole-biotite gneisses. Member 2 is from 100 to 350 m thick in the western and southern parts of the Kalikorva structure and 600–700 m thick in its northeastern part, entirely composing here the antiform core. Member 3 is variable in composition. In some areas, it is entirely composed of garnet amphibolites, whereas in the others, amphibolites enclose abundant bodies of biotite, amphibole-biotite, and, occasionally, garnet-biotite gneisses. In general, proportion of amphibolites in Member 3 varies from 60–65 to 100%. Member 3 is from 300 to 100 m thick in the northwestern and southern parts of the Kalikorva structure and pinches out in its northeastern part. Member 4 is exposed in the northern limb of the antiform, where it is in direct contact along the thrust plane with rocks of the Kovdor nappe overlying the Chupa nappe. The maximal thickness of this member is 150 m. In terms of composition, it is identical to Member 2 and forms with the latter a single unit in areas, where Member 3 pinches out. The integral thickness of supracrustal rocks exposed in the core of the Kalikorva antiform is up to 1300 m. Thickness values are estimated in structural sections for the members under consideration with the accuracy of ±50 m. Naturally, it is not a thickness of original sediments, but the values approximately characterize the role of distinguished stratigraphic units in the supracrustal rock sequence.

According to analysis of the geological map, geological cross sections, and stratigraphic columns, there are two types of sections exposed in the Kalikorva tectonic window (Fig. 3). Amphibolites prevalent in the northwestern part of the structure constitute 75 to 85% of the section. These are rocks of the Kalikorva Complex defined in the course of geological mapping, scale 1: 50000, in 1951 (Fig. 3, column 1). The northeastern part of the structure is mainly composed of kyanite-bearing garnet-biotite and biotite gneisses similar to rocks of the Chupa Complex of the synonymous tectonic nappe (Fig. 3, column 2). These complexes are separated by a zone of transitional rock composition.

METHODS

The supracrustal sequences under consideration are composed of highly metamorphosed and, usually, repeatedly migmatized rocks. Therefore, reconstruction of their origin based on geological (determination of relationships between geological bodies and morphological analysis), petrochemical, and geochemical methods is a necessary step in their study. The origin of metamorphic rocks was interpreted using different discrimination and classification diagrams and statistical methods of multidimensional correlation (Mil'kevich, 1996). Based on distribution of major, trace, and rare earth elements, we defined a weathering degree of rocks and modeled compositions of their provenances. The comparison of trace element contents with standard values characterizing the well-known sedimentation settings (Bhatia, 1983) enabled geodynamic interpretation of the basin evolution.

Analytical methods. Contents of major elements are determined using the X-ray spectral method of silicate analysis (VSEGEI). Concentrations of Rb, Sr, Y, Nb, Ni, Cr, Co, V, and Ba are measured at the X-ray fluorescence analyzer VRA-30 (IPGG RAS) with the relative error below 10%. The method of instrumental neutron-activation analysis (IPGG RAS) was applied to measure REE concentrations. Samples and standards placed in the cadmium filter were radiated by the neutron flow of 1×10^{14} n cm^{-2} s^{-1} during 50 hours. The γ-activity was measured at 4096-channel analyzer using the "ideal" Ge and Ge-(Li) detectors. The relative analytical error for La, Nd, Sm, and Yb is 5–10%, and for Ce, Eu, Tb, and Lu it is 5–15%.

The Sm–Nd isotopic analysis was performed at the IPPG RAS using the method described previously (Kotov *et al.*, 1995). The blank experimental values during measurements are 0.03–0.2 ng for Sm and 0.1–0.5 ng for Nd. The measured $^{143}Nd/^{144}Nd$ ratios are normalized to $^{146}Nd/^{144}Nd = 0.7219$ and $^{143}Nd/^{144}Nd = 0.511860$ in the La Jolla Nd standard. The analytical error for Sm and Nd contents is ±0.5% (2σ), and uncertainty of isotopic ratios $^{147}Sm/^{144}Nd$ and $^{143}Sm/^{144}Nd$ corresponds to ±0.5 and ±0.005%, respectively. The mean weighted $^{143}Nd/^{144}Nd$ value averaged for 11 measurements in the La Jolla Nd standard is equal to 0.511894 ± 8 (2σ). The $\varepsilon_{Nd}(0)$ values and model $T_{Nd}(DM)$ ages are calculated with reference to present-day CHUR values of $^{143}Nd/^{144}Nd = 0.512638$ and $^{147}Sm/^{144}Nd = 0.1967$ after Jacobsen and Wasserburg

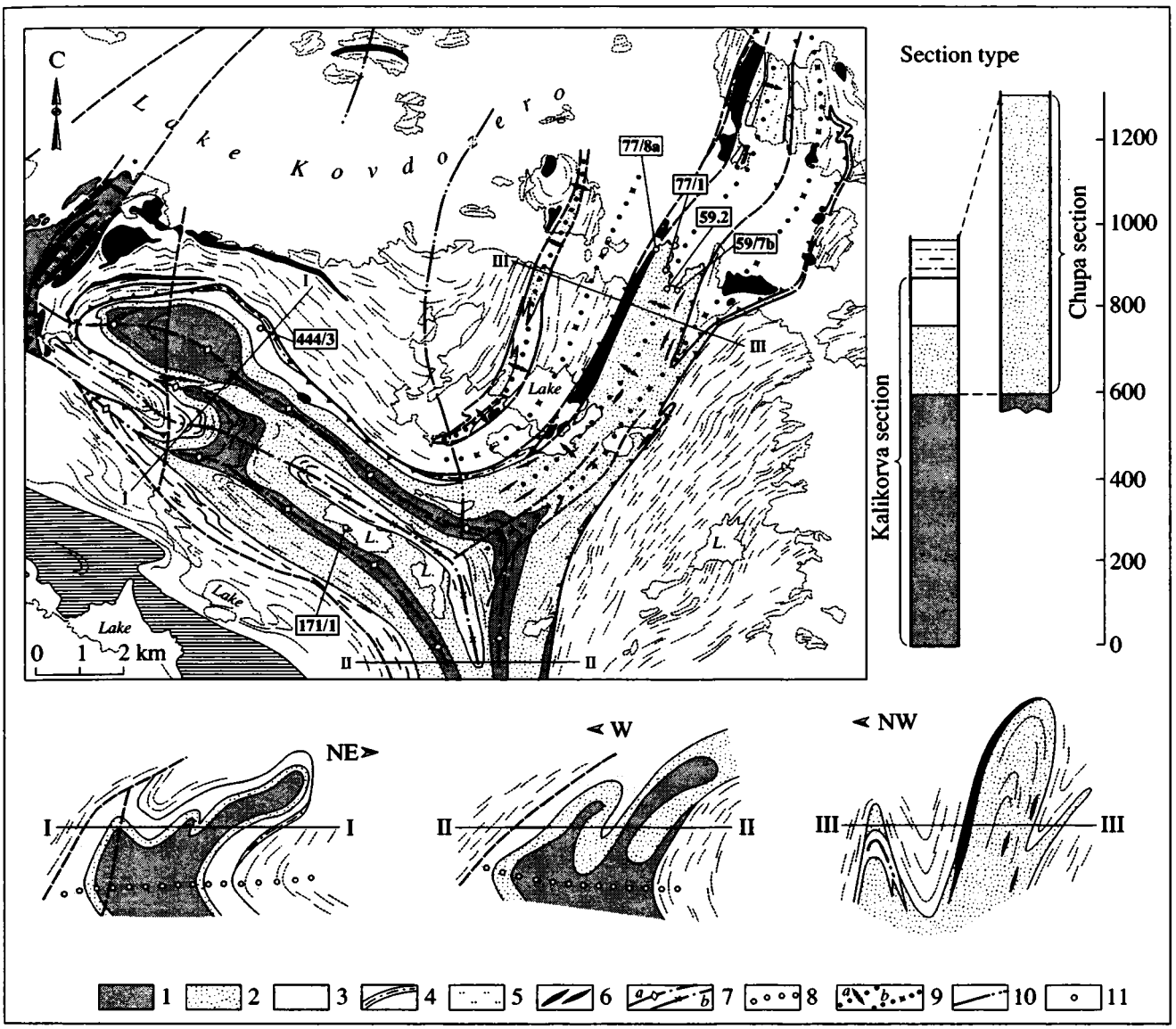

Fig. 3. Geological map of the Kalikorva tectonic window; Chupa nappe: (1) Member 1 of garnet amphibolites (tholeiitic metabasalts) with horizons of biotite and amphibole-biotite gneisses (intermediate and acidic metavolcanics) and with single lenses of biotite-garnet gneisses (metagraywackes), (2) Member 2 of biotite-garnet gneisses with kyanite (metagraywackes) intercalated with horizons of biotite-amphibole gneisses (intermediate and acidic metavolcanics) and amphibolites (metabasalts), (3) Member 3 of garnet amphibolites (tholeiitic metabasalts) with abundant horizons of amphibole-biotite and biotite gneisses (intermediate and acidic metavolcanics), (4) Member 4 of biotite-garnet gneisses with kyanite (metagraywackes), (5) relicts of the Chupa Complex gneisses within younger tonalites; (6) dikes of younger gabbro-norites (Early Paleoproterozoic drusites); folds axes: (7) F_3 antiform (a) and synform (b), (8) F_4, (9) F_5 of antiform (a) and synform (b), (10) F_6; (11) sampling sites for the Sm–Nd isotopic-geochemical analysis and their numbers (outlined); Roman numbers designate geological cross sections.

(1984) and to DM values of $^{143}Nd/^{144}Nd = 0.513151$ and $^{147}Sm/^{144}Nd = 0.2136$ after Goldstein and Jacobsen (1988). When calculating two-stage model ages $T_{Nd}(DM\text{-}2st)$ as described by Liew and Hofman (1988), we used the average-crust ratio $^{147}Sm/^{144}Nd = 0.12$ (Taylor and McLennan, 1985).

RESULTS

The Chupa Complex. The complex is best studied in the stratotype section located in the Lake Verkhnee Pulongskoe area, where it is well exposed and least altered by superimposed processes than in other areas. The dominated garnet-biotite gneisses, frequently with kyanite, associate here with subordinate garnet-amphibolite. Gneisses are intensely migmatized. Non-migmatized varieties constitute not more than 1% of the total section and occur as relicts in its middle part. Partial melting and melt extraction changed the original element ratios in residual gneisses, which are enriched in Al and more mafic than the initial rocks. Therefore,

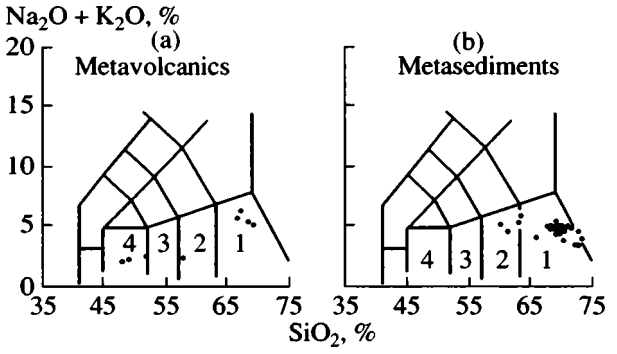

Fig. 4. Classification diagram (Le Bas, 1986) for metavolcanic (a) and metasedimentary (b) rocks of the Chupa Complex; fields in diagram: (1) dacites; (2) andesites; (3) basaltic andesites; (4) low-K basalts.

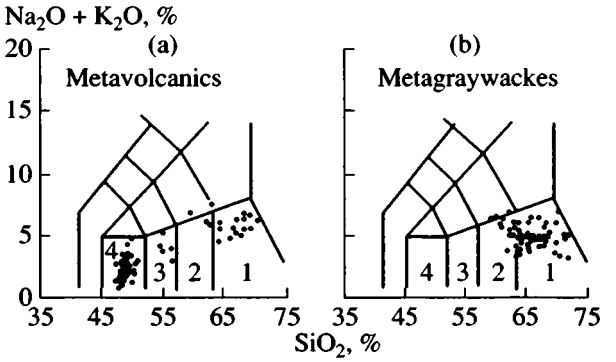

Fig. 5. Classification diagram (Le Bas, 1986) for metavolcanic (a) and metasedimentary (b) rocks of the Kalikorva Complex; fields in the diagram: (1) dacites; (2) andesites; (3) basaltic andesites; (4) low-K basalts.

only non-migmatized rock varieties are taken into consideration by reconstruction of their protoliths.

The gneisses are usually lacking original sedimentary structures, although sometimes they exhibit banding, which may be considered as relict indication of bedding. This feature is usually manifested as alternation of well-expressed melanocratic and leucocratic bands, which are occasionally of the asymmetric structure and resemble sedimentary rhythms.

Chemical compositions of rocks from the Chupa Complex are presented in Table 1 and illustrated in the diagram SiO_2–(Na_2O +K_2O) of Fig. 4. Amphibolites after tholeiitic basalts are close in composition to the Archean basalts TH-1 (Condie, 1983). By chemical composition, the garnet-biotite gneisses can be divided in two groups that likely differ in origin. The first lesser group includes rocks with geochemical features of acidic volcanics. In the classification diagram (Fig. 4a), data points of these rocks plot in the dacite field. Their average chemical composition is as follows: SiO_2 = 67%, Al_2O_3 = 17%, Fe_2O_3 tot = 3.5%, MgO = 1%, and CaO = 4% (Table 1), and the rocks are referred to the calc-alkaline series. The metadacites show differentiated REE spectra; the chondrite-normalized LREE and HREE concentrations range from 40 to 70 and from 3 to 5, respectively. The $(La/Yb)_N$ value is as high as 21.

Gneisses of the second group show high variations of SiO_2 (68 to 77%) and Al_2O_3 (11 to 15%) contents. In occasional more melanocratic interbeds, the SiO_2 content decreases to 60–63%, and that of Al_2O_3 increases to 16–19%. In the classification diagram (Fig. 4b), data points of these gneisses plot predominantly in the dacite field. Nevertheless, characteristic of the rocks are elevated Fe_2O_3tot (up to 6%) and MgO (1.5–3%) contents but low percentage of CaO (2–3%), i.e., the elevated Mg/Ca ratios in the rocks are relatively high. In addition, what is most important, they have high concentrations of Cr (up to 300 ppm and higher) and other elements of the Fe group that is untypical of igneous dacites (Table 1). Such a peculiar composition of gneisses can be explained by sedimentary origin of their protoliths (Myskova *et al.*, 2000). According to classification of Pettijohn *et al.* (1973), the inferable sedimentary protoliths are referred to graywackes.

The $T_{Nd}(DM)$ values estimated for metadacites and metagraywackes of the Chupa Complex vary within narrow limits: from 2.8 to 2.9 Ga (Table 2).

The Kalikorva Complex

The Kalikorva structure defined in the northwestern part of the White Sea region is composed of supracrustal rocks surrounded by granitoids of the tonalite-trondhjemite association (Fig. 3). According to structural data described above, the Kalikorva supracrustal complex is exposed under tonalites of the Karelian craton in a tectonic window.

Analytical results presented in Table 1 and SiO_2–(Na_2O+K_2O) diagram (Fig. 5) imply that amphibolites of the Kalikorva complex correspond to basalts of the tholeiitic series slightly differentiated into Mg- and Fe-enriched rock varieties (Fig. 6). Amphibole-biotite and some biotite gneisses are comparable with intermediate to acidic volcanics (Table 1, Fig. 5a) of the calc-alkaline series (Fig. 6). Dominant are dacites, whereas dacitic and basaltic andesites are subordinate. The main chemical parameters of the rocks are as follows: SiO_2 = 60–70%, MgO = 2–3%, and CaO = 4–6%. In addition, they have high Sr and Ba but and slightly elevated Cr concentrations (125 ppm Cr in average). Metadacites show differentiated chondrite-normalized REE spectra: the normalized LREE and HREE values range from 50 to 100 and from 3 to 10, respectively; the $(La/Yb)_N$ value is 40. Garnet-biotite and biotite gneisses fall in the dacite field of classification diagram (Fig. 5b). They differ however from intermediate and acidic volcanics having significant variations of SiO_2, Al_2O_3, and TiO_2 contents, elevated MgO (>5%) and lowered CaO (2 to 3%) concentrations (Mg/Ca values are elevated consequently). Another important feature is a high concentra-

tion of chromium (up to 400 ppm) and other elements of the Fe group. Because of peculiarities in chemical composition, the gneisses under consideration are different from any known groups of igneous rocks. According to classification of Pettijohn *et al.* (1973), the garnet-biotite gneisses of the Kalikorva Complex correspond to graywackes. Variations in Cr contents discriminate two groups of graywackes (Table 1), the clastic material of which was derived from different model provenances (see discussion below). Metagraywackes with the high Cr contents are prevalent. Characteristic of them are significant variations in the SiO_2 and TiO_2 contents (59–72 and 0.35–0.89%, respectively), the highest MgO (up to 5.8%) and Cr (300–400 ppm) concentrations but lowest CaO (1.7%) contents (Table 1). Less common low-Cr metagraywackes are likely constituents of the lower part of the paragneiss member. Typical of them are relatively low SiO_2 variations (60–68%), lower MgO (up to 4.4%) and Cr (up to 200 ppm, occasionally higher) concentrations but higher CaO (up to 2.6%) content.

The $T_{Nd}(DM)$ age determined for metasedimentary and metavolcanic rocks of the Kalikorva Complex is 2.9 Ga (Table 2).

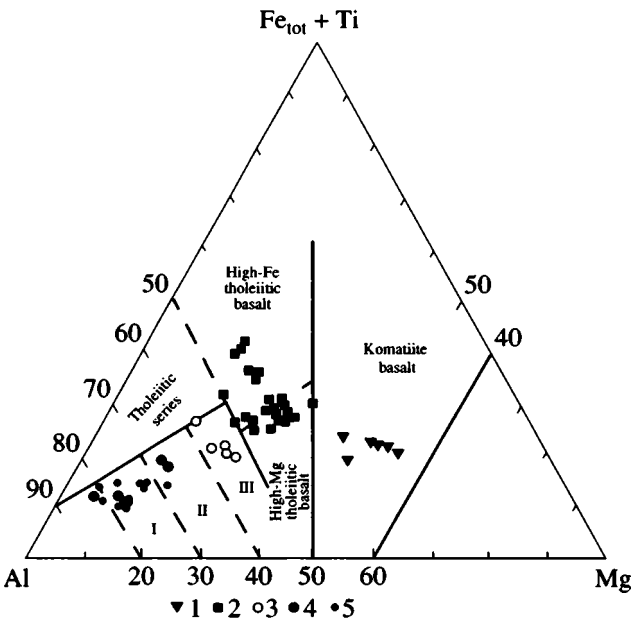

Fig. 6. Metavolcanic rocks of the Kalikorva Complex in the diagram Al-(Fe_{tot}+Ti)-Mg (Jensen, 1976): (1) high-Mg basalt, (2) basalt, (3) basaltic andesite, (4) dacitic andesite, (5) dacite; fields in diagram: (I) dacites, (II) andesites, (III) basalts of the calc-alkaline series.

DISCUSSION

The petrochemical and geochemical peculiarities of graywackes from the Chupa and Kalikorva complexes point to a low provenance maturity and short-distance transport of clastic material. This is also evident from low values of index CIA = 53–55 (Table 1). Inasmuch as the source material experienced no significant alteration during sedimentation, the sediment composition was mainly determined by composition of eroded rocks. The high SiO_2, Na_2O, MgO, FeO, Ti, Cr, Ni, and Co contents in graywackes suggest that both the acidic and mafic rocks were subjected to erosion in the provenance (Tables 3, 4). The acidic material could be derived either from tonalites of the presumable ancient basement or from dacitic lavas associated with graywackes. The mafic material was most likely derived from basalts of the Kalikorva and Chupa complexes or from mafic zones of the Khetolambina nappe. Binary diagrams (Figs. 7, 8) illustrating compositional variations of the Chupa and Kalikorva metagraywackes support this assumption. The behavior of elements during sedimentation is best demonstrated by changes of their concentrations relative to Ti, an element of most inert behavior (Nesbitt, 1979). As is seen in Fig. 7, straight lines connecting in binary diagrams the data points of average metadacite and metabasalt correspond to mechanical mixtures of these rocks. In most diagrams, data points of the Chupa graywackes plot along mixing lines (Fig. 7), thus implying no fractionation of almost all elements in the course of sedimentation, except for Ca, Si, Al, Sr, and Cr. Variable contents of last elements may be explained by influence of sedimentation processes. In case of Ca and Sr, data points plot below the mixing line, which may indicate decomposition of plagioclase and insignificant removal of both elements by chemical weathering. The elevated SiO_2 contents in the SiO_2–TiO_2 plot are related to accumulation of weathering-resistant terrigenous quartz. An "oblique" trend of Al_2O_3 relative to the mixing line is probably a consequence of differentiation during sedimentation (Sawyer, 1986). Position of data points above the mixing line in the Cr–TiO_2 plot is indicative of ultramafic rocks subjected to erosion in the provenance and responsible for the high Cr content in metasediments. In the chondrite-normalized REE diagram (Fig. 9a), REE spectra of the Chupa graywackes occupy intermediate position between dacites and basalts of the Chupa Complex, and the latter rock types could represent a possible source of clastic material for graywackes. Steep LREE segments of the spectra may point to a substantial contribution of dacitic clastic material.

Unlike the Chupa graywackes, their counterparts from the Kalikorva Complex are divisible in two groups (Table 1). Clastic material for rocks of both groups was most likely derived from metavolcanics (dacites and basalts) of the Kalikorva Complex proper. Ultramafic rocks and komatiite basalts also contributed clastic components to graywackes with the highest Cr contents. This is well evident in the Cr–TiO_2 plot (Fig. 8a), where data points of metagraywackes are located above the dacite-basalt mixing line, and also in the binary diagram (Fig. 10), where they are displaced toward the ultramafic end member. The behavior of

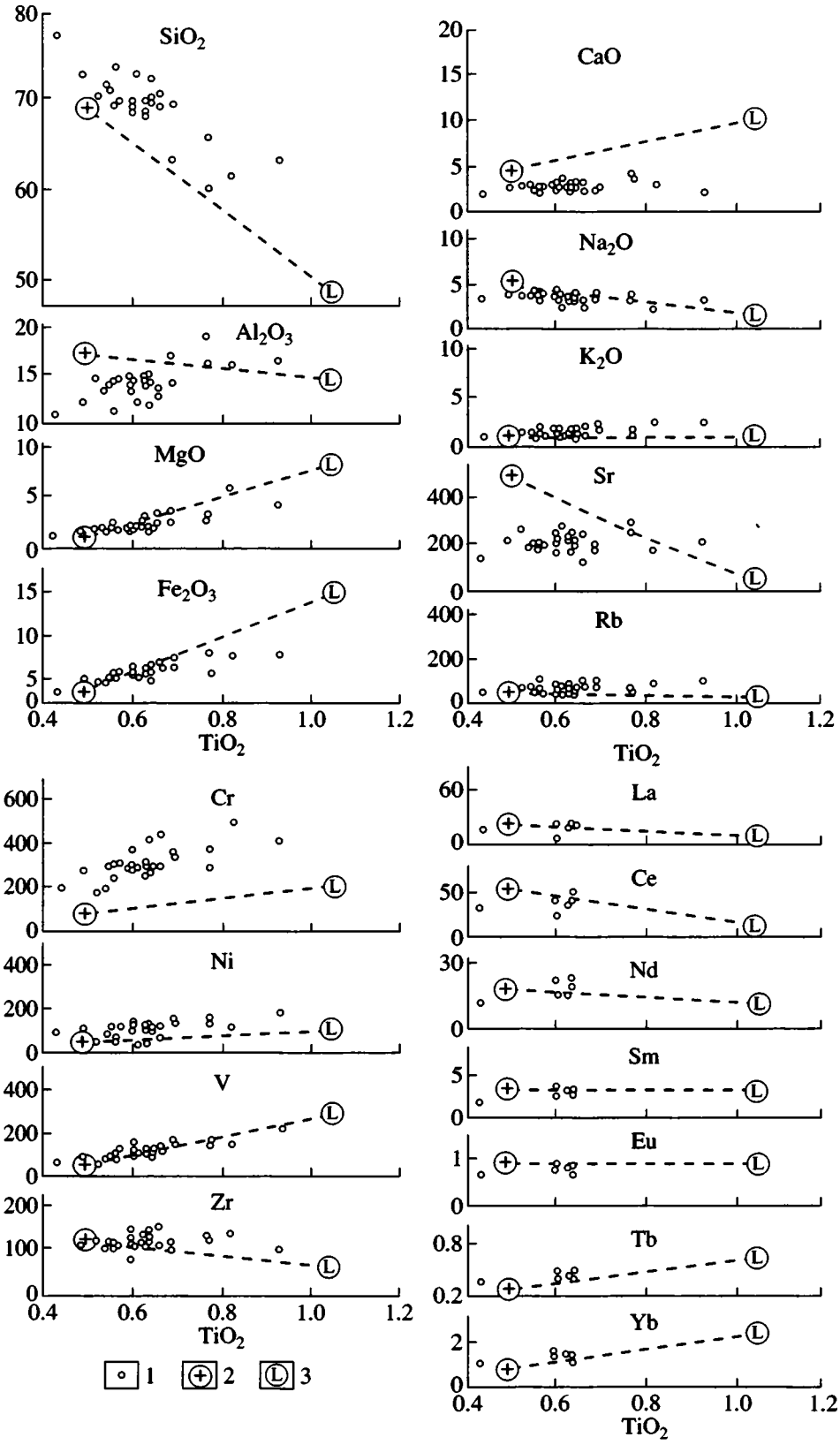

Fig. 7. Variation diagrams for paragneisses of the Chupa Complex: (1) metagraywackes; (2) average compositions of metadacites from the Chupa Complex; (3) average compositions of metabasalts from the Chupa Complex.

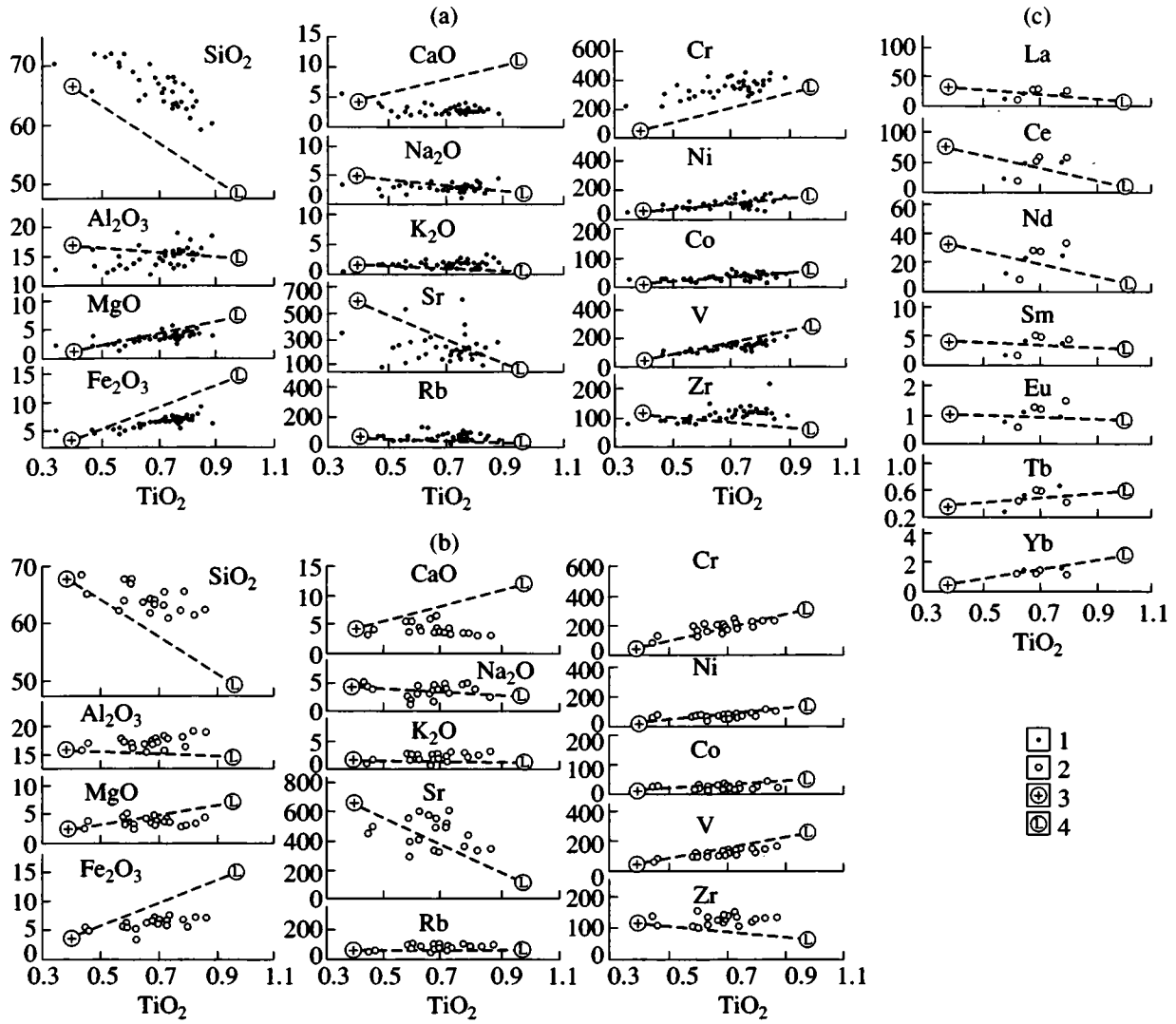

Fig. 8. Variation diagrams for paragneisses of the Kalikorva Complex: (1) metagraywackes with high Cr contents (a); (2) metagraywackes with low Cr contents (b); (3) average compositions of metadacites from the Kalikorva Complex; (4) average compositions of metabasalts from the Kalikorva Complex.

major and trace elements in the Kalikorva metagraywackes characterizes them as immature sediments, whose compositions depended mostly on proportions of rocks in the provenance. Chemical weathering is able to influence mobility of Ca and Sr only; it usually leads to impoverishment of rocks in these elements (Fig. 8). Differentiation in the course of sedimentation results in some enrichment of rocks with SiO_2 and in "oblique" position of Al_2O_3 trends relative to the mixing line. Like in the case of Chupa graywackes, chondrite-normalized REE spectra for these rocks of the Kalikorva Complex occupy an intermediate position between dacitic and basaltic spectra, being displaced toward the former in the LREE segments (Fig. 9b).

Variation diagrams enable discrimination of elements least mobile during sedimentation. These are Ti, Ni, Mg, Fe, and Cr. Since the contents of these elements in sediments depend largely on their concentrations in source rocks, they can be used to estimate the rocks proportions in the provenance. The model compositions of provenances for the Chupa and Kalikorva graywackes are characterized in Tables 3 and 4. As is evident from the tables, graywacke compositions certainly reflect proportions of the rocks mixed. Par example, being most leucocratic, sedimentary rocks of the Chupa Complex imply a significant contribution of acidic material (up to 70%), whereas more melanocratic graywackes of the Kalikorva Complex reveal a higher proportion of mafic components (up to 40%).

In all REE-TiO_2 diagrams (Figs. 7, 8c), data points of the Chupa and Kalikorva metagraywackes plot along or close to the mixing line of source rocks, thus pointing to an insignificant REE fractionation during sedimentation. Accordingly, the Nd model age of meta-

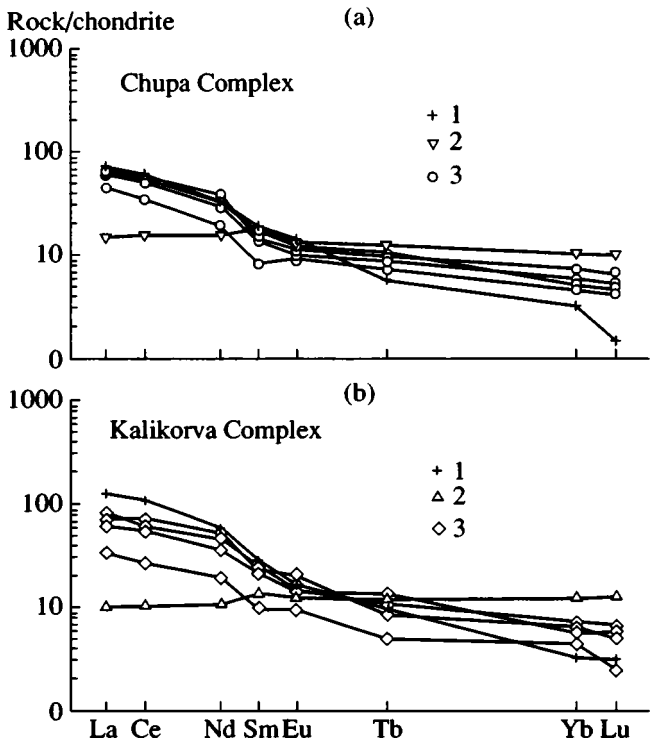

Fig. 9. Chondrite-normalized REE concentrations in metagraywackes and metavolcanics of the Chupa (a) and Kalikorva (b) complexes: (1) dacites; (2) basalts; (3) metagraywackes of the Chupa (a) and Kalikorva (b) complexes.

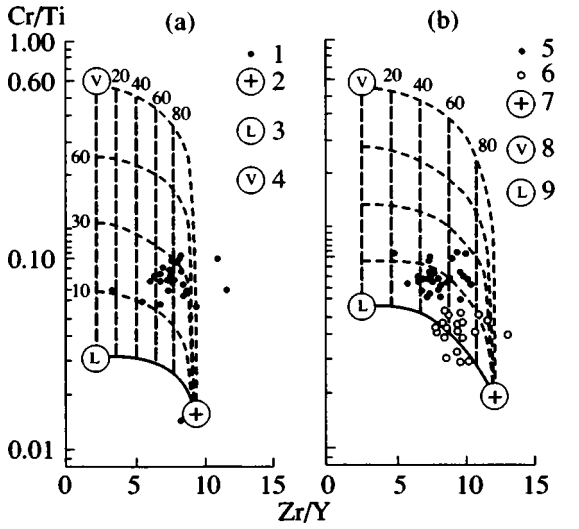

Fig. 10. Geochemical diagram (Camire *et al.*, 1993) illustrating composition of provenance for metaterrigenous rocks of the Chupa (a) and Kalikorva (b) complexes: (1) metagraywackes, (2) metadacites, and (3) metabasalts of the Chupa Complex; (4) metakomatiites of the Kalikorva Complex; (5) high-Cr metagraywackes, (6) low-Cr metagraywackes, (7) metadacites, (8) metabasalts, and (9) metakomatiites of the Kalikorva Complex.

graywackes (Table 2) characterizes the average Nd model age of source rocks. The Nd model age of 2.9 Ga calculated for graywackes of the stratotype area and U-Pb age of their zircons (2.9–3.0 Ga; Bibikova *et al.*, 2001) define the lower age limit of their formation. On the other hand, detrital zircons from northern blocks of the BMB are 3.2 Ga old (Bibikova *et al.*, 1999), and this may be indicative of older rocks occurrence in the provenance. The latter are known in the Karelian craton, but their contribution was most likely insignificant. Otherwise, there would be significant probability to find sediments with a high CIA index, indicative of a long-distance transport of terrigenous material, and with older Nd model ages. Zircons from dacites of the Pulongskoe section are 2870 ± 30 Ma old (Bibikova *et al.*, 2001), and these rocks may be considered as probable main components in the model source of the Chupa graywackes. In addition, similar to metagraywackes, the Nd model age of these dacites is 2.9 Ga and, judging from the $\varepsilon_{Nd}(T)$ value of +3.6, they originated from a juvenile source.

Similar Nd model ages of 2.8–2.9 and 2.9 Ga have been calculated for metagraywackes and metadacites of the Kalikorva Complex, respectively (Table 2). Positive $\varepsilon_{Nd}(2.87)$ values ranging from +3.4 to +4.8 suggest the juvenile origin for dacites.

The metasediments of the Chupa and Kalikorva complexes bear similar geochemical signatures and have close formation ages, which imply a single stratigraphic level of their protoliths accumulation. Based on this important fact, the Chupa and Kalikorva metagraywackes were united into a single complex (Miller *et al.*, 1999). Nevertheless, some peculiarities in composition and spatial position of the rocks are inconsistent with this idea. For instance, the graywacke complexes under consideration are isolated from each other, and their host sections substantially differ in proportions of associated volcanics constituting 5–10% of the Chupa Complex and not less than half of the Kalikorva complex section. It is also important that geochemical characteristics of graywackes point to autonomous provenance. Being similar in contents of major and trace elements, volcanics of the complexes in question show some differences in REE contents. Dacites from the Kalikorva Complex are enriched in LREE as compared with counterparts of the Chupa Complex, whereas contrary relations are characteristic of basalts present in both complexes (Table 1). There are also differences in proportions of rocks mixed in two provenances.

On the other hand, it is likely that sedimentary associations of both complexes formed in similar geodynamic settings. Chemical compositions of studied gneisses suggest that their sedimentary protoliths were composed of almost fresh rock clasts. Sediments of this kind could be formed under condition of a rapid rock disintegration and instant burial of terrigenous material

Table 1. Representative analyses of main rock types from the Chupa and Kalikorva complexes

Component	Chupa Complex						Kalikorva Complex								
	4300/12	4655/5	4654/1	4657/5	4289/2	4296/1	35/9	35/10a	59/2	444/3	77/1	59/7b	171/1	77/8a	47/1
SiO_2	49.00	66.75	69.89	70.16	72.47	77.52	49.07	67.90	67.94	67.68	61.54	70.51	71.83	65.55	64.90
TiO_2	1.04	0.50	0.60	0.64	0.64	0.43	0.99	0.39	0.41	0.41	0.69	0.56	0.58	0.80	0.65
Al_2O_3	14.13	17.84	13.45	13.95	11.83	10.92	14.02	16.49	16.36	15.57	17.00	13.26	12.86	16.01	15.66
Fe_2O_3	14.86	2.76	6.09	6.02	6.56	3.60	14.79	3.26	3.19	3.86	6.39	5.68	5.10	5.17	6.16
MnO	0.20	0.04	0.07	0.08	0.08	0.05	0.23	0.06	0.04	0.06	0.10	0.15	0.07	0.08	0.08
MgO	8.00	1.15	2.18	1.77	1.91	1.40	7.52	1.61	1.74	1.57	4.43	3.10	2.28	2.59	3.58
CaO	10.38	5.03	2.53	2.81	3.04	1.88	11.23	4.01	3.37	5.25	5.88	3.37	2.11	3.03	3.84
Na_2O	1.73	4.72	3.53	3.47	2.73	3.22	1.94	4.26	5.08	4.02	1.47	1.93	3.32	4.63	3.82
K_2O	0.59	1.02	1.60	1.09	0.65	0.93	0.16	1.80	1.77	1.44	2.32	1.35	1.77	2.01	1.20
P_2O_5	0.07	0.18	0.06	0.01	0.07	0.05	0.07	0.22	0.17	0.13	0.19	0.08	0.08	0.14	0.13
Ba	130	478	344	321	67	240	15	822	781	821	692	265	450	378	283
Rb	5	49	72	48	30	40	3	57	43	51	77	47	36	62	48
Sr	65	622	216	205	214	143	114	912	557	215	455	160	311	443	345
Zr	52	119	114	132	150	103	60	110	103	112	120	106	79	137	118
Y	23	14	16	16	20	9	26	8	9	10	14	15	10	14	15
Nb	6	8	6	12	7	<5	<5	<5	<5	<5	<5	5	<5	6	5
Cr	186	44	286	286	412	188	291	53	60	155	138	321	276	182	270
V	297	47	103	133	91	64	304	50	41	50	130	122	91	125	121
Ni	107	18	107	99	120	79	148	23	29	42	40	120	58	55	110
Co	56	13	30	21	29	23	51	<10	<10	12	27	25	12	25	23
CIA			52	53	52	53					52	55	53	51	52
La	4.80	22	21	21	21	15	3.4	43	31		25		11.5	25	21
Ce	13.3	50	42	49	42	30	8.9	97	62.6		53		23.8	62.2	51
Nd	9.91	18	19	23	19	12	6.7	37	28.8		30		12.3	34.1	24
Sm	3.35	3.32	2.81	3.21	2.81	1.66	2.71	5.45	3.79		5.37		1.96	4.68	4.46
Eu	0.97	1.01	0.76	0.90	0.76	0.76	0.98	1.15	0.93		1.26		0.74	1.54	1.09
Tb	0.62	0.28	0.43	0.50	0.43	0.36	0.56	0.47	0.28		0.63		0.24	0.43	0.54
Yb	2.32	0.70	1.3	1.1	1.3	1.0	2.68	0.71	0.49		1.3		0.93	1.36	1.5
Lu	0.34	<0.1	0.18	0.16	0.18	0.14	0.42	0.10	<0.05		0.18		0.08	0.17	0.22
Total REE	35.61	95.36	87.48	98.87	87.48	60.92	26.35	184.88	127.91		116.74		51.55	129.48	103.81

Note: Sampled rock types are (4300/12) metabasalt, (4655/5) metadacite, (4654/1, 4657/5, 4289/2, 4296/1) metagraywackes; (35/9) metabasalt, (35/10a, 59/2, 444/3) metadacites, (77/1, 59/7b, 171/1, 77/8a, 47/1) metagraywackes; major elements in wt % recalculated for dry residue, trace and rare earth elements in ppm; total iron is calculated as Fe_2O_3; according to Nesbitt and Young (1982) CIA = $100[Al_2O_3/(Al_2O_3 + CaO + Na_2O + K_2O]$ (in mol %).

Table 2. Sm–Nd isotopic data for rocks of the Belomorian mobile belt

Sample	Rock	Sm, ppm	Nd, ppm	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ ($\pm 2\sigma$ meas.)	$\varepsilon_{Nd}(0)$	$\varepsilon_{Nd}(T)$	$T_{Nd}(DM)$	$T_{Nd}(DM\text{–}2st)$
Chupa supracrustal complex									
4655/5	Metadacite	2.83	17.13	0.1000	0.510989	–32.2	3.6	2883	2885
4654/1	Metagraywacke	3.42	18.53	0.1114	0.511206	–27.9	3.6	2882	2883
4296/1	Metagraywacke	1.96	12.36	0.0958	0.510888	–34.1	3.2	2910	2920
4657/5	Metagraywacke	3.211	18.851	0.1029	0.511009	–31.8	2.9	2931	2943
Kalikorva supracrustal complex									
59/2	Metadacite	3.77	26.69	0.0853	0.510771	–36.4	4.8	2811	2790
444/3	Metadacite	3.29	19.86	0.1000	0.510978	–32.4	3.4	2897	2903
77/1	Metagraywacke	4.49	25.43	0.1066	0.511121	–29.6	3.7	2875	2875
59/7b	Metagraywacke	2.82	15.33	0.1112	0.511189	–28.3	3.4	2901	2904
171/1	Metagraywacke	2.72	15.38	0.1068	0.511090	–30.2	3.0	2923	2931
77/8a	Metagraywacke	4.94	29.09	0.1027	0.511055	–30.9	3.9	2862	2860

Note: $\varepsilon_{Nd}(t)$ and $T_{Nd}(DM\text{-}2st)$ are calculated for the age of 2870 Ma.

in proximity to the provenance. There is some discrepancy between model proportions of rocks in provenances and their real abundance in the studied sections. Two alternatives are possible: either there were deeply eroded volcanic structures already transformed into terrigenous material, or the eruption products of closely spaced volcanic edifices experienced instant disintegration. Interesting in this connection are data on formation settings of metagraywackes from the English River paragness belt (Superior Province, Canada). Van de Kamp and Beakhouse (1979) who studied these rocks suggest that pyroclasts erupted by underwater volca-

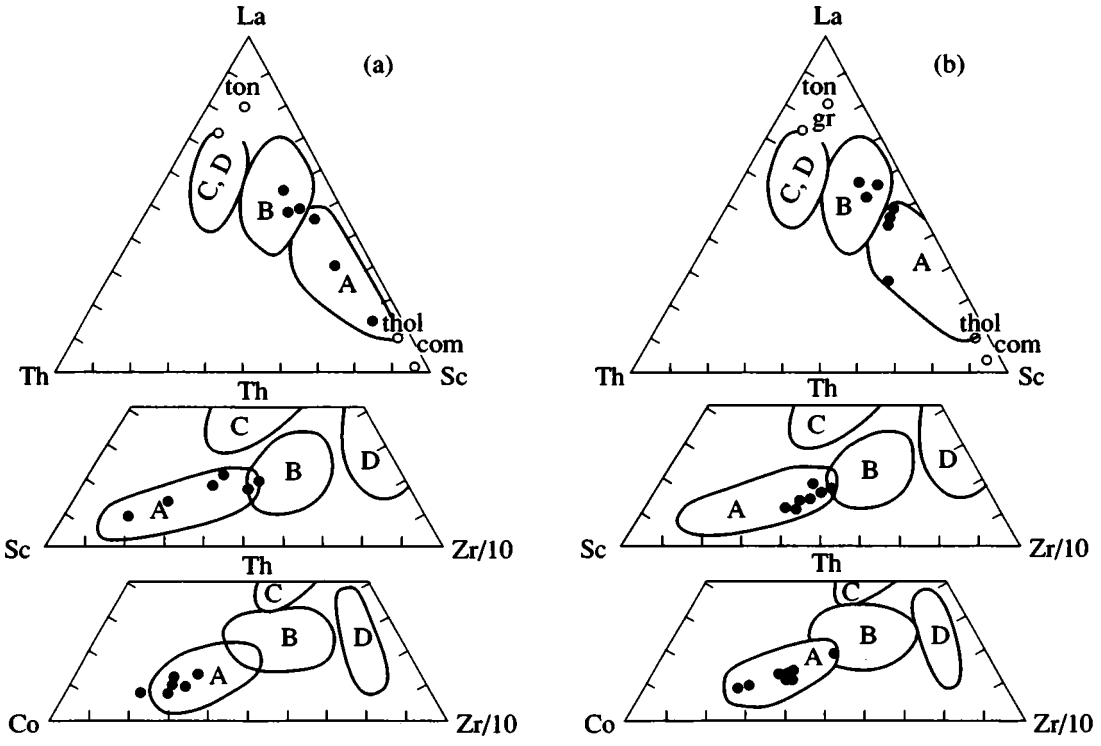

Fig. 11. Discrimination diagrams (Bhatia, 1983) for metagraywackes of the Chupa (a) and Kalikorva (b) complexes; fields in diagrams: (A) oceanic island arcs; (B) continental island arcs; (C) active continental margins; (D) passive continental margins; (gr) granite; (ton) tonalite; (com) komatiite; (thol) tholeiite.

Table 3. Model composition of provenance for metagraywackes of the Chupa Complex (major elements in wt %, trace elements in ppm)

Component	Average contents of components in metavolcanics and their contribution to sediments						Model concentrations	Average contents of components in metagraywackes
	dacite		basalt		komatiite			
	average contents	contribution (70%)	average contents	contribution (20%)	average contents	contribution (10%)		
TiO_2	0.50	0.35	1.04	0.2	0.58	0.06	0.61	0.62
Ni	37	26	107	20	463	46	92	87
Fe_2O_3	3.5	2.4	14.86	2.9	13.99	1.4	6.7	5.65
MgO	1.06	0.7	8	1.6	14.48	1.4	3.7	2.08
Cr	80	56	186	37	1718	172	265	289

Table 4. Model composition of provenance for metagraywackes of the Kalikorva Complex (major elements in wt %, trace elements in ppm)

Component	Average contents of components in metavolcanics and their contribution to sediments						Model concentrations	Average contents of components in metagraywackes with high Cr concentrations
	dacite		basalt		komatiite			
	average contents	contribution (60%)	average contents	contribution (30%)	average contents	contribution (10%)		
TiO_2	0.41	0.25	0.96	0.29	0.58	0.06	0.6	0.67
Ni	25	15	125	38	463	46	99	100
Fe_2O_3	3.7	2.2	14.43	4.2	13.99	1.4	7.8	6.51
MgO	1.77	1.0	7.74	2.3	14.48	1.4	4.7	3.62
Cr	63	38	313	102	1718	172	312	336

Component	Average contents of components in metavolcanics and their contribution to sediments				Model concentrations	Average contents of components in metagraywackes with high Cr concentrations
	dacite		basalt			
	average contents	contribution (60%)	average contents	contribution (40%)		
TiO_2	0.41	0.25	0.96	0.38	0.63	0.66
Ni	25	15	125	50	65	56
Fe_2O_3	3.7	2.2	14.43	5.7	7.9	5.7
MgO	1.77	1.0	7.74	3.0	4.0	3.18
Cr	63	38	313	125	163	177

noes represented the primary clastic material of graywacke sedimentary basin. This explains also an insignificant chemical weathering of that material. Paragneisses of the Tertiary Tokiwa Formation in Japan represent similar rocks (Fiske and Matsuda, 1964).

The available isotopic-geochemical and geochronological data (Bibikova *et al.*, 1999) imply that supracrustal rocks of the Chupa and Kalikorva complexes are the oldest ones in the BMB. Rocks with older crustal isotopic signatures have not been found among tonalites surrounding these supracrustal complexes. The Nd model age calculated for tonalites (2.8–2.9 Ga; Bibikova *et al.*, 1999) is comparable with the age estimated for supracrustal rocks, and it is impossible to assume therefore that the tonalite-trondhjemite rock association was widespread in denudation areas. Only an insignificant admixture of clastic material derived from ancient sources of the Karelian craton is admissible. Judging from peculiarities of rocks, which most likely represented the basement (oceanic tholeiitic basalts) and simultaneously contributed material, along with dacites of the calc-alkaline series, to formation of metagraywackes, protoliths of the latter could be accumulated in settings surrounded by immature island arcs

(*Geodinamicheskie rekonstruktsii*, 1989). The analysis of composition of metasedimentary rocks in discrimination diagrams (Bhatia, 1983), where the Chupa and Kalikorva metagraywackes fall into the field of oceanic island arcs (Fig. 11), leads to the same inference.

CONCLUSION

The available geological, isotopic-geochemical, and geochronological data allow the following conclusions to be drawn.

(1) The examined sections of supracrustal rocks are exposed in the single Chupa nappe that originated at the early (subduction) stage of development of the mobile belt plunging under the Karelian craton.

(2) The U–Pb dating of metamorphic zircons (Bibikova *et al.*, 2001) and the Sm–Nd isotopic-geochemical study of metasedimentary and metavolcanic rocks combined with geological data show that two types of Chupa nappe sections can be considered as almost coeval. The most probable time span of their formation ranges from 2.9 to 2.8 Ga. Judging from these data, the rocks of the Keret greenstone belt and volcanics from the examined complexes, as well as presumable proto-ophiolites of mafic zone, but not tonalite-trondhjemites of the Khetolambina and Chupa nappes, could be in provenance of clastic material for graywacke sequences. The tonalite-trondhjemite protolith 2.7–2.8 Ga old is younger than supracrustal rocks (Lobach-Zhuchenko *et al.*, 1993; Bibikova *et al.*, 1999), and crustal history of the latter was short like that of volcanogenic-sedimentary sequences.

(3) Many compositional peculiarities of paragneisses indicate that the structure of examined sequences was determined by local volcanic and sedimentation events responsible for the established difference between the Kalikorva and Chupa complexes as well. Common of both are an extremely low degree of chemical differentiation during sedimentation, a significant mechanical mixing of materials derived from closely-spaced sources, and substantial role of mafic, mostly volcanogenic material. All these features lead to the conclusion that supracrustal rocks formed in areas surrounded by juvenile oceanic (ensimatic) island arcs.

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Microfossils from the Upper Riphean Type Section (the Karatau Group of Southern Urals)

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Abstract—New materials studied (over 200 samples) and published data critically analyzed are used to specify composition of organic-walled microfossils and to elucidate their vertical and lateral distribution in the Upper Riphean type section, the Karatau Group of southern Urals that is divided into six formations. New burials of microfossils are discovered in the Zilmerdak, Inzer, Min'yar, and Uk formations. Microfossils most important in terms of stratigraphy are established in a basal interval of the group (Nugush and Bederysh subformations of the Zilmerdak Formation) and near the base of the Inzer Formation. According to results obtained, the microfossil assemblages from above subdivisions include some transit taxa formerly unknown and abundant morphotypes of a complex structure, which appear for the first time at these stratigraphic levels. The pre-Nugush, pre-Bederysh, and pre-Inzer levels are most significant among microphytological boundaries established in the Karatau succession of microfossils. The first of them is marked by the first occurrence of large acritarchs *Chuaria* and *Navifusa* in association with morphotypes of filamentous and other complex structures. The particular species of acanthomorphic *Trachyhystrichosphaera* and associated *Arctacellularia, Tortunema*, and *Glomovertella* forms are characteristic of the second level, whereas the third one signifies appearance of *Prolatoforma, Tasmanites, Pellicularia, Eosaccharomyces*, and some other taxa. The comparative analysis of South Uralian and other regional successions of microfossils implies that the microbiotas are applicable for biostratigraphic subdivision of the Upper Riphean.

Key words: Riphean stratotype, biostratigraphy, microfossils, microphytological levels, siliciclastic successions.

INTRODUCTION

Microfossils cannot be used in the Riphean biostratigraphy without the reliable information on their distribution in the type and reference sections of this age. An objective integral model of the ancient microbiota evolution must be based on the analysis of natural successions of Riphean microbiotas and their homotaxial fragments and on the reconstruction of ecological structure of microfossil communities in paleobasins of that time (Veis *et al.*, 2001). In recent years, investigations of this kind were carried out in several regions of Siberia, where the discrete Riphean microbiotas and their representative successions have been studied in detail in the aspects of taxonomy and facies-ecology (Veis *et al.*, 1998a, 1998b, 1999; Nagovitsin, 2001; Sergeev and Lee Seong-Joo, 2001; Veis and Vorob'eva, 2002), while microphytological characterization of the Riphean stratotype is less well understood. Meanwhile, according to the chronostratigraphic concept, the age of a section bearing microfossils can be correctly understood only after its correlation with the Riphean stratotype, which must be substantiated by all available methods (Semikhatov, 1995, 1997). To be significant in terms of biostratigraphy, ages of concrete microbiotas from regional successions can be defined therewith based on their positions in the vertical succession of microbiotas in the Riphean of southern Urals. That is why there was a new wave of interest to the advanced microphytological characterization of the Riphean stratotype (Veis and Larionov, 1999; Veis *et al.*, 2000; Sergeev and Lee Seong-Joo, in press).

The Karatau Group is of prime significance from the standpoint of aforementioned problems. This group, the terminal one in the Riphean type succession and representing stratotype of the Upper Riphean erathem, is quite complete in terms of stratigraphy, composed of slightly altered rocks, and hosting the majority of microfossil burials established during recent years in southern Urals (Veis *et al.*, 1990; Kozlov *et al.*, 1991, 1994; Mikhailova and Podkovyrov, 1992; Sergeev, 1992; Maslov *et al.*, 1994, 2001). In addition, several important microbiotas of presumably Late Riphean age, those with morphotypes of a complex structure included, are known to occur within and nearby the Upper Riphean stratotype area (Kozlova, 1987; *Mikrofossilii...*, 1989; Veis and Larionov, 1999 and others).

The biostratigraphic analysis of Riphean microfossils from the Urals has been beyond the research attention for nearly twenty years (Yankauskas, 1980a, 1980b, 1982; Keller and Yankauskas, 1980; *Stratotip*

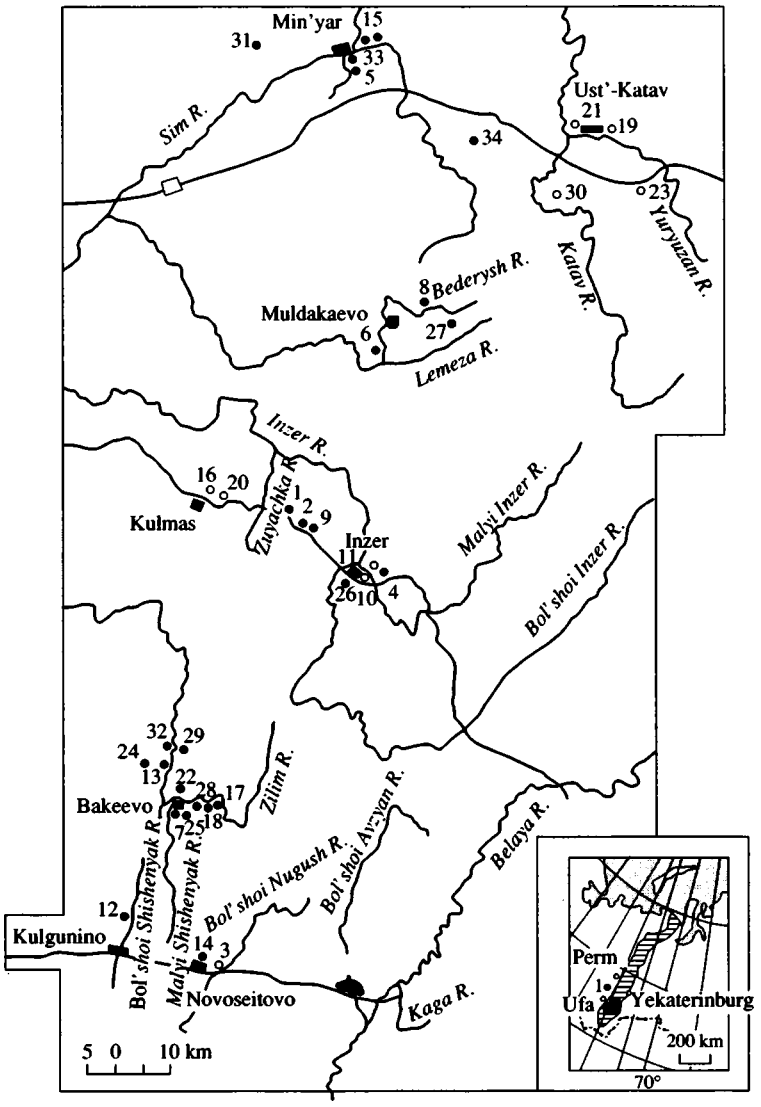


Fig. 1. Geographic localities of sampled sections of the Karatau and Asha Groups of the Upper Precambrian, the Bashkirian meganticlinorium of southern Urals: (1) Biryan, (2, 3, 23) Nugush, (4) Lemeza, (5–9 and 22, 24) Bederysh subformations of the Zilmerdak Formation; (10, 11, 30) Katav Formation; (12, 31) lower and (13–15, 25–27, 32–34) upper Inzer subformations; (15) lower Min'yar Subformation; (16–19, 28, 29) lower Uk subformation, all of the Karatau group; (20, 21) Zigan Formation of the Asha Group (filled and open circles denote sites of fossiliferous and barren samples, geographic and stratigraphic positions of samples are discussed in the text).

rifeya..., 1982). Our work is aimed to cover this lacuna using new materials (over 200 samples), which characterize the best sections of the Karatau Group (Fig. 1) and nearby rock successions of a problematic stratigraphic status, and to summarize the published data. The main objectives are as follows: (1) to broaden the paleontological characterization of the group; (2) to analyze stratigraphic ranges of formerly known and newly detected microfossils and to coordinate them with the currently adopted structure of the group; (3) to assess objectively a possibility of microphytological subdivision of the group and the ages of correlative microbiotas known in other areas of southern Urals and Cisuralian region. The succession of Riphean subdivisions used as a basis is that described in summarizing works (*Stratotip rifeya...*, 1983; *Unifitsirovannye...*, 1994).

GEOLOGY OF STUDIED SECTIONS

The Karatau Group that yielded the analyzed microfossils is a thick (3.0–5.3 km) sedimentary succession representing, along with the Burzyan, Yurmata, and Kuzha groups, the latter discriminated in places, the Riphean type section in the western flank of southern Urals. Sediments of the group are widespread in the western limb of the vast (110 × 350 km) Bashkirian meganticlinorium of a debatable structure. Some

researchers (Senchenko, 1976; and others) believe that the meganticlinorium represents a system of large autochthonous anticlinal and synclinal folds, while the others (Kamaletdinov, 1974, 1977; and other works) distinguish here a series of allochthonous thrust sheets in a subparallel arrangement. According to the third opinion (Ivanov and Ivanov, 1997; Vinogradov *et al.*, 2000, 2001), the structure in question includes large allochthonous blocks of the Earth's crust amalgamated in the course of tectogenesis. Two former viewpoints are better substantiated and imply that the Burzyan-Karatau succession is representative, characterizing the Riphean Eonothem, which is one of major subdivisions in the Late Precambrian chronostratigraphic scale of Russia (Semikhatov *et al.*, 1991). In the latter hypothesis, the Burzyan, Yurmata (+ Kuzha), and Karatau groups are regarded in contrast as subdivisions, which are sub-synchronous, corresponding to synonymous Riphean erathems, or unsatisfying the natural stratigraphic succession, and accordingly the scale validity is put under doubts. In our opinion, the indicated large lithostratigraphic subdivisions represent the vertical succession, and the Riphean chronostratigraphic scale is considered in this work as a valid one.

Inside the Bashkirian meganticlinorium, the distinctly eroded surface of the Karatau Group is overlain by Vendian deposits. In most sections, the group is distinguished above the Yurmata Group, but relationships between these two subdivisions have been interpreted variably. As has been alternatively suggested, (1) the groups are interrelated via transitional strata as a rule; (2) they are separated everywhere by a considerable hiatus with associated structural transformations; (3) in places, they are separated by the Kuzha Group that spans, to a considerable extent, the Yurmata-Karatau hiatus and belongs to the Upper Riphean (see Filippov, 1991, and Veis *et al.*, 2000, for a detailed discussion).

In the regional stratigraphic schemes accepted at present (*Unifitsirovannye...*, 1993), the Karatau Group includes six subdivisions: (1) the Zilmerdak Formation of siliciclastic deposits (1950–2950 m); (2) the Katav Formation of variegated thin-banded clayey limestones and marls (200–400 m); (3) the Inzer Formation ranging in composition from fine-grained glauconite sandstones to shales and stromatolitic limestones (350–680 m); (4) the Min'yar Formation predominantly composed of dolostones (500–600 m); (5) the Uk Formation of carbonate-terrigenous deposits (180–400 m); and (6) the Krivaya Luka Formation of sandstones and shales (400–500 m), which are preserved after the pre-Vendian erosion in the eastern limb of the Bashkirian meganticlinorium only. The last subdivision has been formerly regarded as correlative with the Tolpar Formation of sandstones (650 m) and with the Suir Formation (300 m) of shales intercalated with tilloids, both attributed now to the Vendian. Boundaries between the Zilmerdak, Katav, and Inzer formations are transitional in contrast to basal levels of the Min'yar, Uk, and Krivaya Luka formations, which reveal distinct erosion marks of variable significance (*Stratotip rifeya...*, 1983; Raaben, 2001). Recent data of C-isotope chemostratigraphy suggest that in the upper interval of the group there is a series of hiatuses, importance of which was underestimated before (Podkovyrov *et al.*, 1998).

In most sections, the Zilmerdak Formation is divisible into the Biryan (1400–2000 m), Nugush (200–300 m), Lemeza (100–250 m), and Bederysh (250–400 m) subformations. Sandstones of variable grain size and composition are dominant rocks of the first and third subdivisions, whereas siltstones and shales are characteristic of two other subformations. Limestones and marls of the Katav Formation are divided into the lower red (100–250 m) and upper variegated (100–150 m) subformations. The structure of two Inzer subformations is dissimilar in eastern and western areas of the study region. In eastern sections exposed within the Inzer synclinorium, both Inzer subformations (550–550 and 250–300 m) are composed of alternating packages of sandstone-shale composition, and boundary between them is conventional. In western sections, the lower (Podinzer) subformation (100–300 m) includes the basal Iken Member of shales and dark limestone beds of variable thickness (0 to 90 m), whereas the upper one is composed of two terrigenous successions (70 and 50 m) separated by a limestone member (80 m). In the Min'yar Formation, the lower Minka Subformation of limestones and dolostones (200–300 m) is discriminated from the upper Byanka Subformation of cherty dolostones (250–350 m). The Uk Formation includes the lower subformation of sandstones and shales (70–150 m) and the upper one of clayey limestones (up to 300 m). The Magadeevo (300 m), Apshak (90 m), and Peresheikovskii (120 m) subformations of the Krivaya Luka Formation differ from each other in proportions of sandstone beds (*Stratotip rifeya...*, 1983; Kozlov *et al.*, 1991, 1994, 1997; Maslov, 1997 and others).

The lithologic composition and peculiar succession of indicated formations and subformations have been established at the early investigation stages of particular sections and whole distribution area of the Karatavian deposits in southern Urals (*Stratigraficheskii...*, 1994, and references therein). However, principal problems of lateral relationships between terrigenous and carbonate stromatolite-bearing subdivisions, which were under heated discussion for a long time (e.g., Raaben, 1975, 1981, 1985; Krylov, 1983), have been partially or completely solved in the last decade only (Kozlov *et al.*, 1991, 1994; Maslov and Krupenin, 1991; *Unifitsirovannye...*, 1993; Maslov, 2002).

Formerly, the distribution area of Karatavian deposits in southern Urals used to be divided into the inner and outer facies zones (*Stratotip rifeya...*, 1983), but now it is a common practice to recognize the eastern and central Karatau Group type sections in the former zone and the western sections in the latter zone (Kozlov, 1982; Kozlov *et al.*, 1994). In the inner zone, the widespread outcrops of Upper Riphean deposits are

concentrated in the extended Inzer synclinorium and in a narrow belt outside the latter, which is situated along the eastern limb of the Yamantau anticlinorium. In the outer zone, these deposits are sporadically exposed in a series of comparatively smaller structures, such as the Karatau, Adzhigardak, Yalmash-Uryuk, and other anticlines. Distinctions between the Karatavian sections of the inner and outer zones are most clearly evident within the interval above the Zilmerdak Formation, since proportions of terrigenous sediments in that interval are much greater in the east that in the west (Kozlov, 1982; *Stratotip rifeya...*, 1983; Maslov, 1997).

Depositional environments of the Karatau Group have been controversially interpreted. According to sedimentological reconstructions done by Maslov (1988, 1997), sediments of the Zilmerdak formation accumulated in diverse continental (fluvial-deltaic), supershallow, shallow coastal, and shelf settings, whereas deposits of the Katav-Uk interval correspond predominantly to marine subtidal facies. In addition to the Biryan and Lemeza shoaling trends indicative of the Karatavian basin retreat from the study region, he distinguished within the Zilmerdak interval the Nugush and Bederysh episodes of the basin deepening, when sediments accumulated in a shelf zone below the storm wave base, i.e., under quiet hydrodynamic environments. In opinion of Maslov, the overlying sediments recorded a gradual transition from shallow marine settings of the Katav-Inzer time to a deeper open basin of the Min'yar time, which experienced shoaling in the Uk time. Analyzing sedimentogenesis in the Middle-Upper Karatavian interval, other researchers came to another conclusion and suggested that the post-Zilmerdak succession was deposited under conditions of more frequent sea-level oscillations in a relatively shallow sea basin. In the course of these oscillations, the lower Katav, Podinzer, and lower Min'yar deposits predominantly accumulated in deeper settings near the storm wave base, whereas stromatolitic and clastic limestones, as well as dolostones in upper portions of these and Uk formations, were deposited in much shallower shelf zones with active hydrodynamic environments (Podkovyrov *et al.*, 1998).

Isotopic age determinations for the Karatavian sediments of southern Urals have been obtained using different methods (Podkovyrov *et al.*, 1998; Ovchinnikova *et al.*, 2000; and references therein). First, these are the K–Ar dates for mineralogically unstudied glauconites from a series of successive horizons of the group. They represent the following upward-decreasing series of values: 938 Ma for the upper Katav Subformation, 896–853 Ma for the lower-middle interval of the Inzer Formation, 791–683 Ma for the upper member of the latter, and 740 and 713–680 Ma for the lower and upper Min'yar subformations, respectively. Recent and more reliable isotopic-geochronological information about the Karatau Group deposits is as follows. The Pb–Pb isochron date of 836 ± 25 Ma has been obtained for the lower Inzer (Sim) limestones and interpreted as elucidating age of the early diagenesis (Ovchinnikova *et al.*, 1998). The Rb–Sr isochron date of 835–805 Ma determined for the authigenic illite from the Inzer Formation is assumed to define the age of burial diagenesis of this subdivision (Ovchinnikova *et al.*, 1995). The weighted average Pb–Pb age of 780 ± 85 Ma is considered as approximating the formation time of the upper Min'yar dolostones (Ovchinnikova *et al.*, 2000). The concordant K–Ar and Rb–Sr dates equal to 688 ± 10 and 670 ± 10 Ma, respectively, have been determined for Al-glauconites from the lower Uk Subformation (Gorozhanin and Kutyavin, 1986).

In addition, the Rb–Sr dating of whole-rock samples from the Zilmerdak, Inzer, and Min'yar formations showed that clayey sediments of these Karatavian subdivisions of southern Urals experienced epigenetic transformations 740 ± 20, 800 ± 40, and 800 ± 40 Ma ago, respectively. However, the same method used to determine age of epigenetic alterations in the Kuzha and Yurmata groups yielded unexpected results. The date of 525 ± 30 Ma that has been obtained for the latter group is much younger that the value of 835 ± 70 Ma estimated for the former one and comparable with that characterizing the Karatau Group (Vinogradov *et al.*, 2000, 2001; Vinogradov and Gorozhanin, 2002).

It has been inferred, based on the above dates, that the lower and upper age limits of the Karatau Group correspond to 1000 ± 50 and 650 ± 20 Ma, respectively (Semikhatov *et al.*, 1991; Kozlov *et al.*, 1991, 1994, 1997; Podkovyrov *et al.*, 1998; Raaben, 2001). The corresponding time span is somewhat narrower than the interval from 1030 ± 20 to 600 ± 10 Ma that has been estimated for the Late Riphean based on a synthesis of original and published results (Semikhatov, 2000; Semikhatov *et al.*, 2000). We should mention therewith that the age calibration of any general chronostratigraphic unit outside the stratotype area of the latter cannot be regarded as unambiguous, if correlation of a given subdivision with the stratotype is not substantiated by all possible methods, the microphytological one included. In this aspect, it is especially important that there are cardinally different opinions with respect to the Late Riphean age of some Siberian successions perfectly studied in terms of paleontology.

Recent data of C-isotope chemostratigraphy (Podkovyrov *et al.*, 1998) showed that positive $\delta^{13}C$ values predominantly characteristic of Karatavian carbonate deposits have the variation amplitudes growing upward within the Katav-Uk interval in a good accord with the general variation trend of this parameter in Neoproterozoic deposits of some other regions (Kaufman and Knoll, 1995; Knoll *et al.*, 1995). In addition to this similarity, Podkovyrov and his colleagues emphasized a difference in configurations of the South Uralian and model variation curves and a lower limit of $\delta^{13}C$ values in the upper interval of the Late Riphean type section, which admitted the aforementioned assumption that characteristic of that interval are considerable strati-

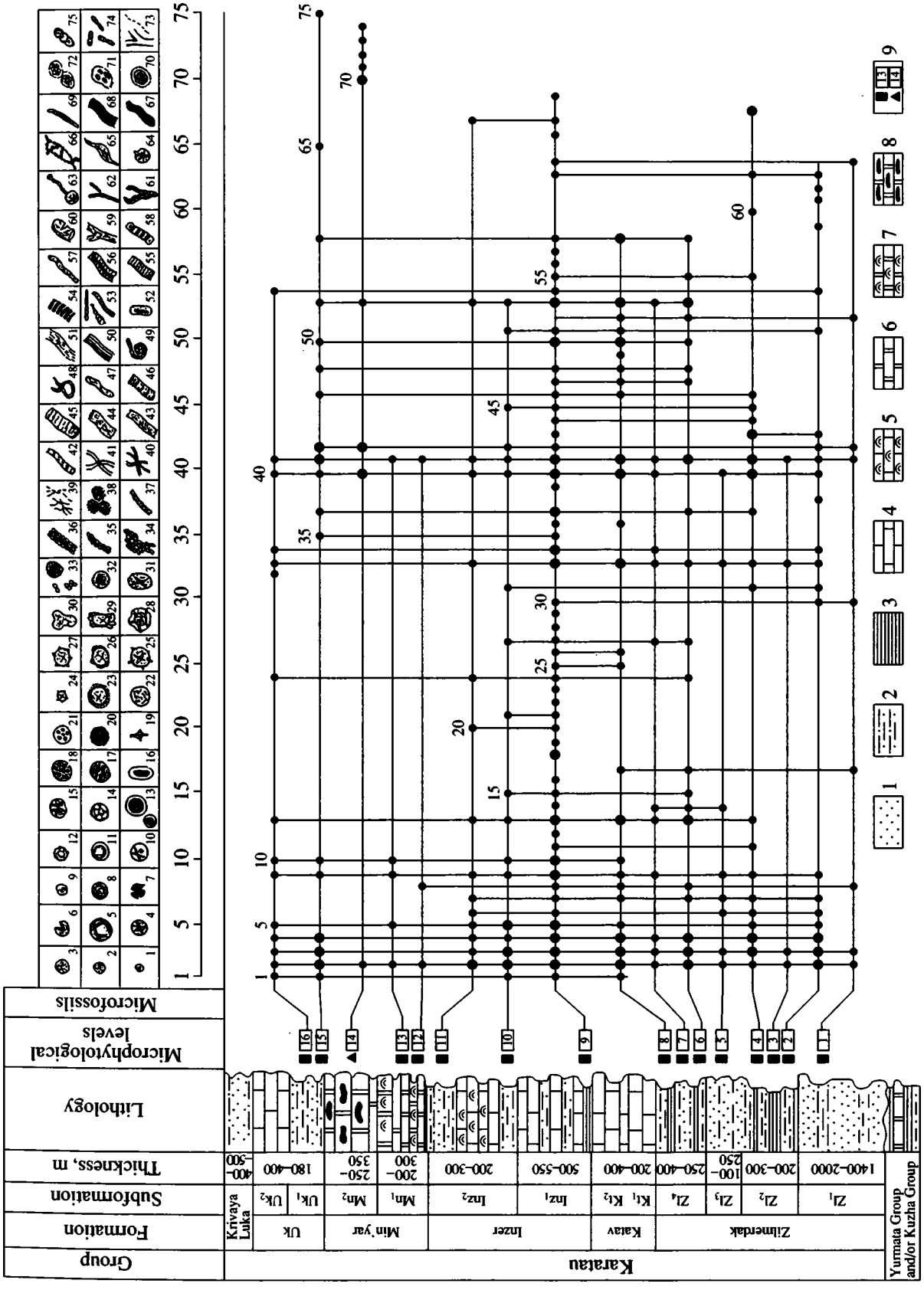
graphic hiatuses (Podkovyrov *et al.*, 1998). Moreover, the comparative analysis of in-phase oscillations within initial segments of both curves led to the conclusion, along with biostratigraphic, geochronological, and geological data, that the model curve characterizes nearly the entire Karatavian type section rather than its second half (Raaben, 2001).

Microphytological data considered below characterize mostly the Karatavian deposits of southern Urals within limits of the Bashkirian meganticlinorium. The only exception is an upper part of representative Riphean section with abundant microfossils, which has been recovered by Borehole Kabakovo-62 in a peripheral area of the Russian platform 28 km southeastward of Ufa near the western boundary of the Uralian foredeep. Paleontological remains from this site are of especial significance, because they have been originally integrated into a general trend of stratigraphic distribution of Riphean microbiotas in the southern Urals and used to elaborate the regional biostratigraphic schemes for this subdivision of the Riphean (Keller and Yankauskas, 1980; Yankauskas, 1980, 1982; *Mikrofossilii...*, 1989). In that borehole, microfossils of the Karatavian affinity have been found mainly in the siliciclastic Tukaevo and Ol'khovyi formations, which are attributed to the middle and Upper Riphean, and at some levels of the overlying Priyutovo (Karmaly) and Shikhan (Savaleevo) formations of carbonate terrigenous sediments, which are of a complex inner structure (Keller and Yankauskas, 1980; Andreev *et al.*, 1981; *Stratotip rifeya...*, 1982, 1983; Maslov and Isherskaya, 1998; *Stratigraficheskaya...*, 2000; Romanov and Isherskaya, 2001).

Analyzing taxonomic composition of Karatavian microbiotas, we distinguish, as before (Veis and Vorob'eva, 1993; Veis *et al.*, 1998, 1999), the discrete microphytological levels, i.e., the comparatively thin members with abundant microfossils, which are separated by thicker stratigraphic intervals nearly or completely barren of microfossils. The distinguished levels are numbered successively throughout the Karatau Group section. We do not include into this succession those assemblages of microfossils, which are of a doubtful stratigraphic status, though important for microphytological characterization of the Upper Riphean of southern Urals. These are the Tyul'ma, Kuzha, Tolpar, and Suir microbiotas found within the Riphean stratotype area and the aforementioned Tukaevo, Ol'khovyi, Priyutovo and Shikhan microbiotas from Borehole Kabakovo-62 drilled outside that area. Microbiotas of the first group have been properly analyzed before (Mikhailova and Podkovyrov, 1992; Veis *et al.*, 2000) and are used in this work as a comparative material without revision of their taxonomic composition. Those of the second group are briefly considered after characterization of microfossil assemblages from the Karatavian type section, which are of a close taxonomic composition.

The maceration method and classification of organic-walled microfossils used in this study are identical to those described earlier (Veis and Vorob'eva, 1993, 2002). The studied morphotypes are found in a

Fig. 2. Stratigraphic distribution of organic-walled and mineralized microfossils in the Karatau Group deposits, the Upper Riphean of southern Urals: (1) sandstone and/or conglomerate; (2) siltstone; (3) shale; (4) limestone; (5) stromatolitic limestone; (6) dolostone; (7) stromatolitic dolostone; (8) dolostone with cherty lentils and nodules; (9) microphytological levels of organic-walled (black quadrangles) and mineralized (black triangles) microfossils. Established morphotypes by numbers: (1) *Leiominuscula minuta* (Naum.) (sheaths <10 μm); (2) *Leiosphaeridia crassa* (Naum.), emend. Jank. (sheaths <70 μm); (3) *L. excultpa* (Tim.), emend. Mikh. (sheaths 70–150 μm); (4) *L. jacutica* (Tim.), emend. Mikh. et Jank. (sheaths 150–300 μm); (5) *Chuaria circularis* Wale., emend. Vidal, Ford (sheaths > 300 μm); (6) *Leiosphaeridia bicrura* Jank., sheaths with median breaks; (7) *L. ternata* (Tim.), emend. Mikh. et Jank., sheaths with peripheral breaks; (8) *L. kulgunica* Jank., sheaths with a rimmed aperture (?pylome); (9) *Nucellosphaeridium minutum* Tim. (external sheath <100 μm); (10) *N. nordium* (Tim.), external sheath >100 μm; (11) *Simia nerjenica* A. Weiss; (12) *S. simica* (Jank.); (13) *Pterospermopsimorpha* spp.; (14) *Satka* spp.; (15) *Kirbia multipartite* Mikh. et Volk.; (16) *Konderia elliptica* A. Weiss; (17) *Valeria lophostriata* Jank.; (18) *Tasmanites ripheicus* Jank.; (19) *Octaedrixium* sp.; (20) *Spumosina rubiginosa* (Andr.); (21) *Aimia* spp.; (22) *Cerebrosphaera* sp.; (23) *Cymatiosphaeroides* sp.; (24) *Trachyhystrichosphaera parva* Mikh.; (25) *T. aimika* Herm.; (26) *T. stricta* Herm.; (27) *T. truncata* Herm. et Jank.; (28) *T. vidalii* Knoll; (29) *Prolatoforma aculeata* Mikh.; (30) large acritarchs of irregular shape (monstrous sheaths); (31) *Navifusa majensis* Pjat.; (32) *Tetrasphaera antiqua* Tim. et Herm.; (33) *Myxococcoides* spp.; (34) *Ostiana microcystis* Herm.; (35) *Sphaerocongregus variabilis* Moorm.; (36) *Chlorogloeopsis zairensis* Maithy; (37) *Polysphaeroides contextus* Herm.; (38) *Coniunctiophycus totticus* A. Weiss; (39) *Archaeotrichion contortum* Schopf and/or *A. lacerum* Herm.; (40) *Eomycetopsis robusta* Schopf, *emend.* Knoll et Golubic; (41) *Leiotrichoides* spp. and/or *Heliconema uralense* Jank.; (42) *Siphonophycus* spp.; (43) *Asperatofilum experatus* (Herm.); (44) *Taenitrichoides jaryshevicus* Assejeva; (45) *Plicatidium latum* Jank.; (46) *Rectia costata* (Jank.); (47) *Brevitrichoides bashkiricus* Jank.; (48) *Tortunema sibirica* Herm., emend. Herm.; (49) *Glomovertella* spp.; (50) *Polytrichoides lineatus* Herm., emend. Herm.; (51) *Eomicrocoleus crassus* Horod. et Donald.; (52) *Archaeoellipsoides* spp.; (53) narrow and medium-sized oscillatorian trichomes (ex gr. *Oscillatoriopsis, Filiconstrictosus, Cephalophytarion, Calyptothrix*, etc.; (54) unnamed wide oscillatorian trichomes (>50 μm); (55) *Botuobia magna* (Tynni et Donner); (56) *Palaeolyngbya sphaerocephala* Herm. et Pylina; (57) *Trachytrichoides ovalis* Herm.; (58) *Arctacellularia* spp.; (59) branching thallomes with a weak longitudinal striation ex gr. *Pseudodendron*; (60) *Lakhandinia* sp.; (61) *Majaphyton antiquam* Tim. et Herm.; (62) thalluses of elongated cells (segments) ex gr. *Proterocladus*; (63) *Caudosphaera expansa* Tim. et Herm.; (64) *Germinosphaera* spp.; (65) *Pellicularia tenera* Jank.; (66) *Eosaccharomyces ramosus* Herm.; (67) *Fabiformis porosus* Pjat.; (68) wide compact filaments (thalluses); (69) *Nelcanica curta* Pjat.; (70) *Gloeodiniopsis lamellose* Schopf, *emend.* Knoll, Golubic; (71) *Eogloeocapsa* sp., *Eoaphanocapsa oparinii* Nyberg et Schopf, *Sphaerophycus medius* Horod. et Donald.; (72) *Eoentophysalis cumulus* Knoll et Golubic; (73) *Ramivaginalis uralensis* Nyberg et Schopf and *Biocatenoides* sp.; (74) *Eosynechococcus amadeus* Knoll et Golubic; (75) *Oscillatoriopsis zilimica* (Jank.).


Group
Formation
Subformation
Thickness, m
Lithology
Microphytological levels
Microfossils
Karatau
Yurmata Group and/or Kuzha Group
Zilmerdak
Katav
Inzer
Min'yar
Uk
Kriyaya Luka
Zl1
Zl2
Zl3
Zl4
Kt1
Kt2
In1
In2
Mn1
Mn2
Uk1
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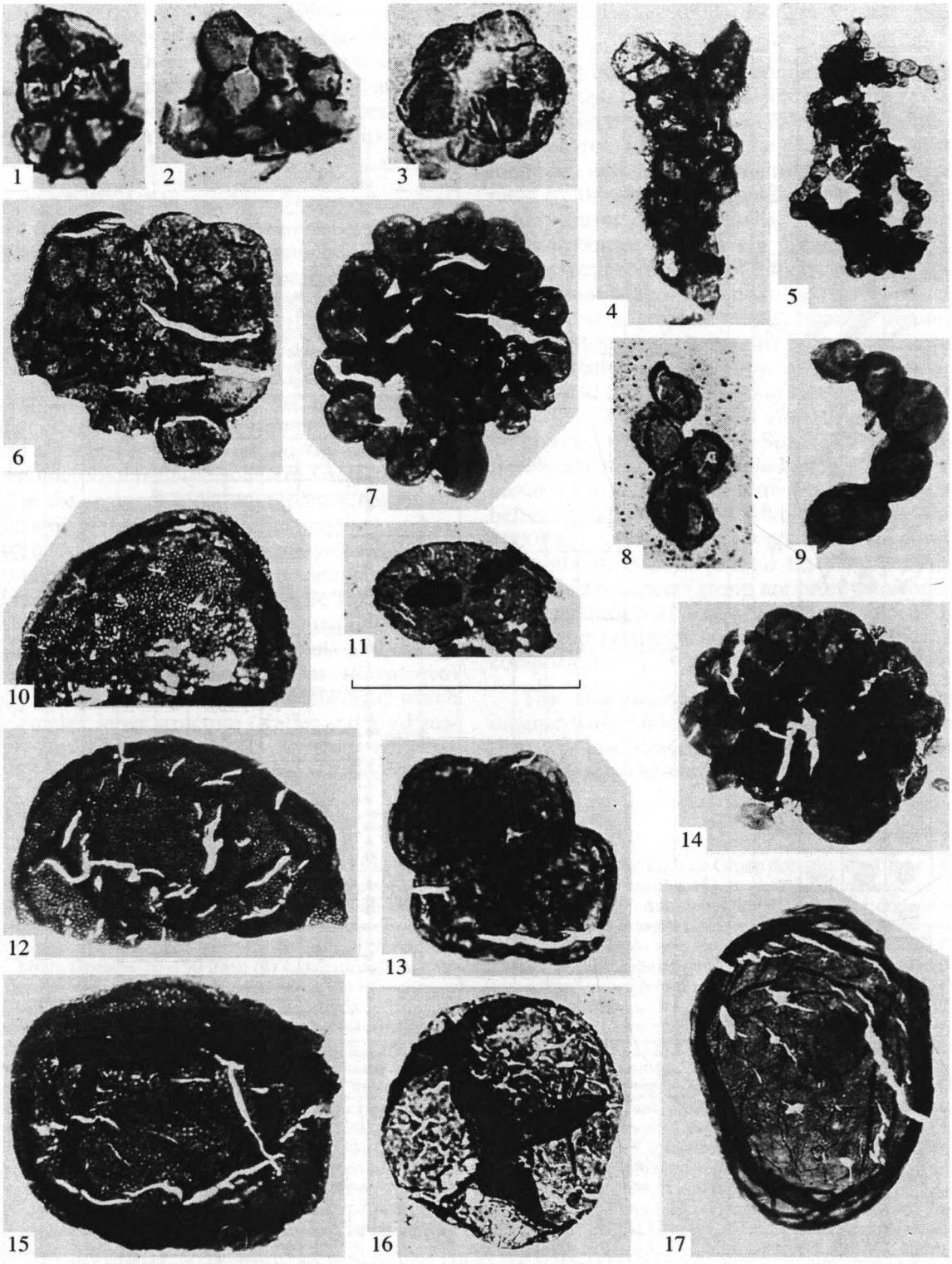
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satisfactory preservation state. They are represented by fragmented or intact organic remains of dark orange to brown coloration, structures of which are variable. The general taxonomic composition of the Karatavian microbiota is illustrated in Fig. 2, and most characteristic morphotypes are pictured in Plates I–III.

MICROFOSSILS FROM THE KARATAU GROUP

Zilmerdak Formation

Biryan, Nugush, and Lemeza subformations (microphytological levels 1–5). The Nugush Subformation (three microphytological levels) of dark siltstones and shales is the only one promising abundance of organic-walled microfossils, while the Biryan and Lemeza subformations (one microphytological level per each) are predominantly composed of light-colored medium- to coarse-grained sandstones nearly barren of organic remains. Owing to their position in the basal part of the Karatavian stratotype, which is significant for local and interregional correlations, all three subformations are of particular interest and have been studied in many localities of southern Urals (Kozlov *et al.*, 1991, 1994, 1997, and references therein). Despite the long-lasted study, the entire pre-Bederysh interval of the Zilmerdak formation was lacking a comprehensive microphytological characterization until recently, and abundant organic-walled microfossils were found here not long ago predominantly in the middle Nugush Subformation within a very limited area (Kozlova, 1990; Kozlov *et al.*, 1994; Veis *et al.*, 2000). Since these microfossils have been analyzed in detail by Veis *et al.* (2000) they are characterized below in a brief manner.

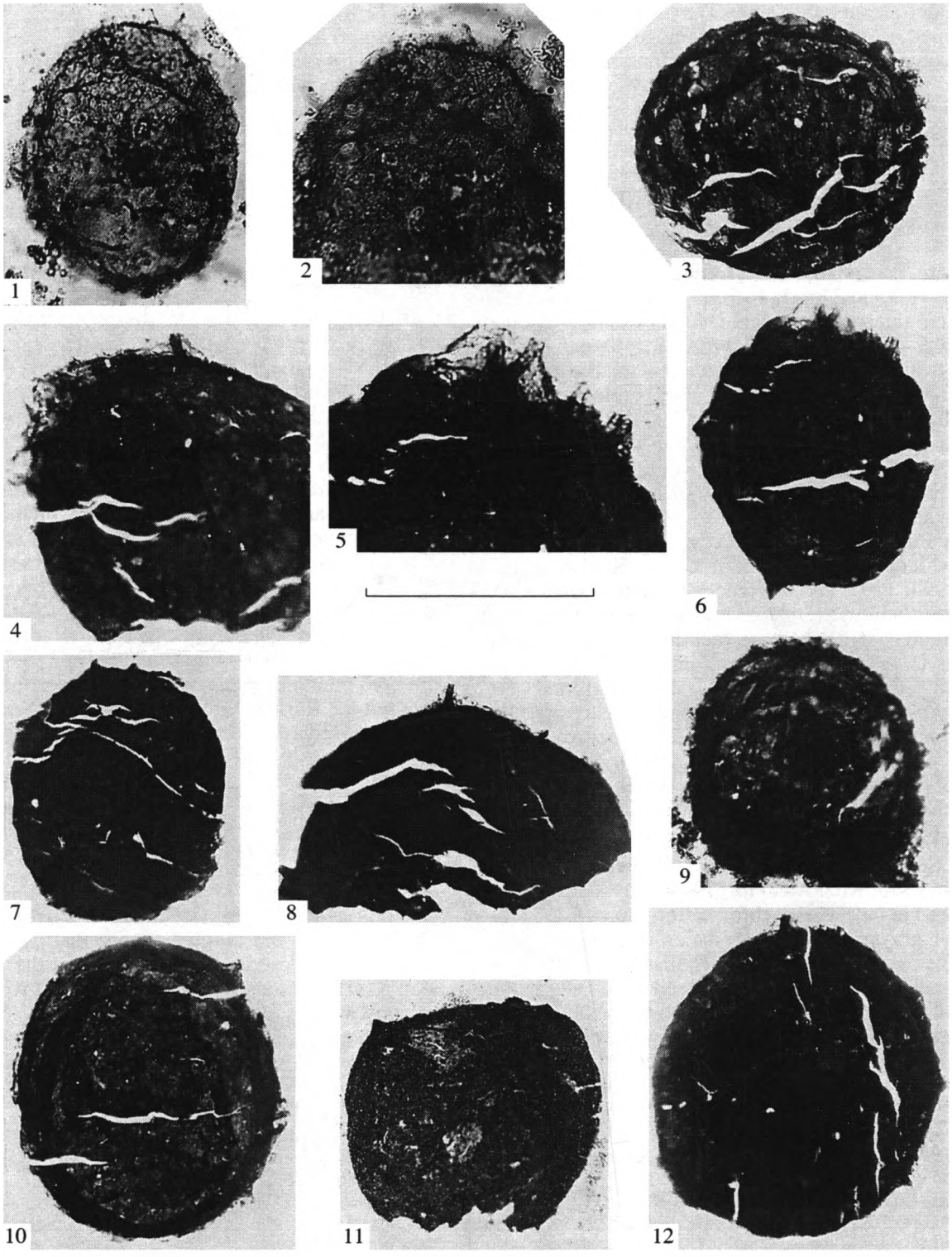
Level 1. All not abundant microfossils from the Biryan Subformation, which are known at present (Kozlova 1990; Kozlova *et al.*, 1994), have been found 200 to 250 m below the subformation top in thin interlayers of gray siltstones and shales intercalated with dominant massive sandstones. Their single locality is established in several closely-spaced road cuts along the Ufa-Inzer highway running close to the Zilmerdak Ridge crest (Fig. 1, point 1) along the western boundary of the Inzer synclinorium. The outcrops yield here small *Leiosphaeridia* (=*Stictosphaeridium*, *Favososphaeridium*) forms, *L. ternata* (Tim.) included, in association with *Satka, Germinosphaera, Myxococcoides, Eomycetopsis, Leiotrichoides*, and small isolated barrel-shaped vesicles attributed by Kozlova to *Arctacellularia*, although they likely belong to akinete-like *Archaeoellipsoides* morphotypes. Samples we collected from variegated siltstones of the Biryan Subformation at the same and close stratigraphic levels turned out to be barren.

Outside the Upper Riphean stratotype area, microfossils presumably of the Biryan level (Kabakovo microbiota from synonymous horizon of the Karatavian) were detected by Yankauskas (see in Andreev *et al.*, 1981) in the Tukaevo and Ol'khovyi (Akberdy) formations penetrated by Borehole Kabakovo-62 in the depth intervals of 5100–5099 and 4765–4762 m, respectively. Microfossil assemblages characteristic of both intervals are identical, consisting of very small *Leiominuscula*, medium-sized *Leiosphaeridia, L. ternata* (Tim.) and *L. kulgunica* (Jank.) included, and *Satka* specimens, occurring in association with large (up to 450 μm) lophostriate *Valeria* similar in size to the Late Riphean *Chuaria* and with fragments of some transit forms (Keller and Yankauskas, 1980; Andreev *et al.*, 1981; *Stratotip rifeya...*, 1982). According to an alternative opinion, organic remains from the lower or from both intervals of the Kabakovo-62 section characterize not the Karatau Group, but the basal Kuzha beds below it, which are locally distinguished in southern Urals either as filling a considerable pre-Karatavian hiatus (Filippov, 1997, 2000, 2001; Veis *et al.*, 2000), or as occupying a lower stratigraphic position and corresponding to the Yurmata Group (Maslov and Isherskaya, 1998; Romanov and Isherskaya, 2001, and references therein).

Levels 2–4. Much more abundant microfossils are characteristic of the Nugush Subformation (Kozlova *et al.*, 1994) exposed not far away from the previous site in road cuts along the same highway 11 to 11.5 km northwestward of the Inzer Settlement (Fig. 1, point 2). The subsequent revision of this locality (Veis *et al.*, 2000) showed that microfossils are confined here to

Plate I. Microfossils of the Upper Riphean Karatau Group, Southern Urals (all pictured specimens are stored at the Geological Institute, Moscow)

(1) *Satka elongata* Jank., specimen no. 994/312; (2–4, 6–9, 14) *Myxococcoides* spp., specimen nos. 994/321 (2), 972/146 (3), 972/405 (4), 972/406 (6), 927/414 (7), 994/323 (8), 972/803 (9), and 972/110 (14) representing shapeless (2), spheroidal (3, 6, 7, 14), *Pleurocapsa*-type (4) or sublinear (8, 9) accumulations of spherical or mutually compressed cells; (5) *Ostiana microcystis* Herm., specimen no. 972/129; (10, 12, 15) *Tasmanites ripheicus* Jank., specimen nos. 972/189, 972/801 (10), 972/801 (12), and 972/103 (15) of large thick-walled envelopes regularly perforated by pores of insignificantly variable size; (11) *Konderia elliptica* A. Weiss, specimen no. 972/181, an ellipsoidal morphotype with internal body copying the shape of outer envelope; (13) *Leiosphaeridia crassa* (Naum.) emend Jank., specimen no. 972/113, an aggregate of four thickened envelopes of identical size; (16) *Cerebrosphaera* sp., specimen no. 972/183, a large morphotype with a characteristic cerebriform envelope; (17) *Nucellosphaeridium nordium* (Tim.), specimen no. 972/605. Specimens by localities: (1, 2, 8) Zilmerdak Formation, Bederysh Subformation, microphytological level 7, Zilim River basin, left bank of the Malyi Shishenyak River 2 km westward of the Bakeevo Village; (3–7, 9–17) Inzer Formation, lower Subformation, microphytological level 9, Zilim River basin, right bank of the Bol'shoi Shishenyak River 6 km northward of the Kulgunino Village. Scale bar sizes: (1, 6) 70 μm; (2) 50 μm; (3, 4, 8, 9) 80 μm; (5, 11) 320 μm; (7, 10, 12, 14) 160 μm; (13) 100 μm; (15) 110 μm; (16) 180 μm; (17) 140 μm.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three levels within the 50-m-thick interval located 130 to 180 m above the subformation base, where dark-colored siltstones and shales are abundant. The least representative assemblage of small *Leiosphaeridia*, *Myxococcoides*, *Leiotrichoides*, and large thallomes with a transverse ornamentation, which are especially remarkable among microfossils, characterizes the third level. Microfossil assemblages of the second and fourth levels are more abundant and diverse and reveal association of morphotypes characteristic of the third level along with large *Leiosphaeridia*, *Nucellosphaeridium*, *Chuaria*, *Navifusa*, and *Asperatofilum*. In addition to these taxa in common, the second level bears subsphaerical colonies of coccoidal *Coniunctiophycus*, thallomes *Ostiana*, filamentous aggregates of *Eomycetopsis* sheaths, tubular ornamented *Siphonophycus*, very broad trichome-like *Caudosphaera*, trilobate thalluses *Majaphyton*, and filaments presumably representing ancient cyanobacteria *Pseudodendron* and green algae *Proterocladus* of the *Cladophora* type. Specific of the fourth level are *Simia*, *Pterospermopsimorpha*, small coccoidal *Polysphaeroides*, wide ribbon-like *Taenitrichoides*, *Plicatidium*, *Rectia*, septate *Botuobia*, and elongated-ellipsoidal *Lakhandinia* characteristic only of this level within the Nugush Subformation. Numerous samples collected from the Nugush Subformation in other sites (Bol'shoi Nugush River basin and Kamennye Gory Ridge; Fig 1, points 3 and 23) did not yield the identifiable microfossils.

Level 5. A single microbiota locality in the Lemeza Subformation is discovered 2 km eastward of the Inzer Settlement (Fig. 1, point 4). Microfossils occur here in sandy siltstones near the subformation top. Like in the Biryan Subformation, their assemblage of a low diversity consists of not abundant *Leiosphaeridia*, *Nucellosphaeridium*, Satka, *Myxococcoides*, and *Eomycetopsis* forms (Veis *et al.*, 2000).

Bederysh Subformation (levels 6–8). Microfossils from this subformation are detected in a series of exposures within the outer (Alatau anticlinorium, Adzhigardak anticline) and inner (Inzer synclinorium) facies zones of the Karatau Group deposits (Mikhailova and Podkovyrov, 1992, and references therein). In addition, Yankauskas (see Andreev *et al.*, 1981) reported on two assemblages of dissimilar composition from different depth levels (3639–3636 and 3528–3526 m) of the borehole section Kabakovo-62, which are presumably attributable to the Bederysh level.

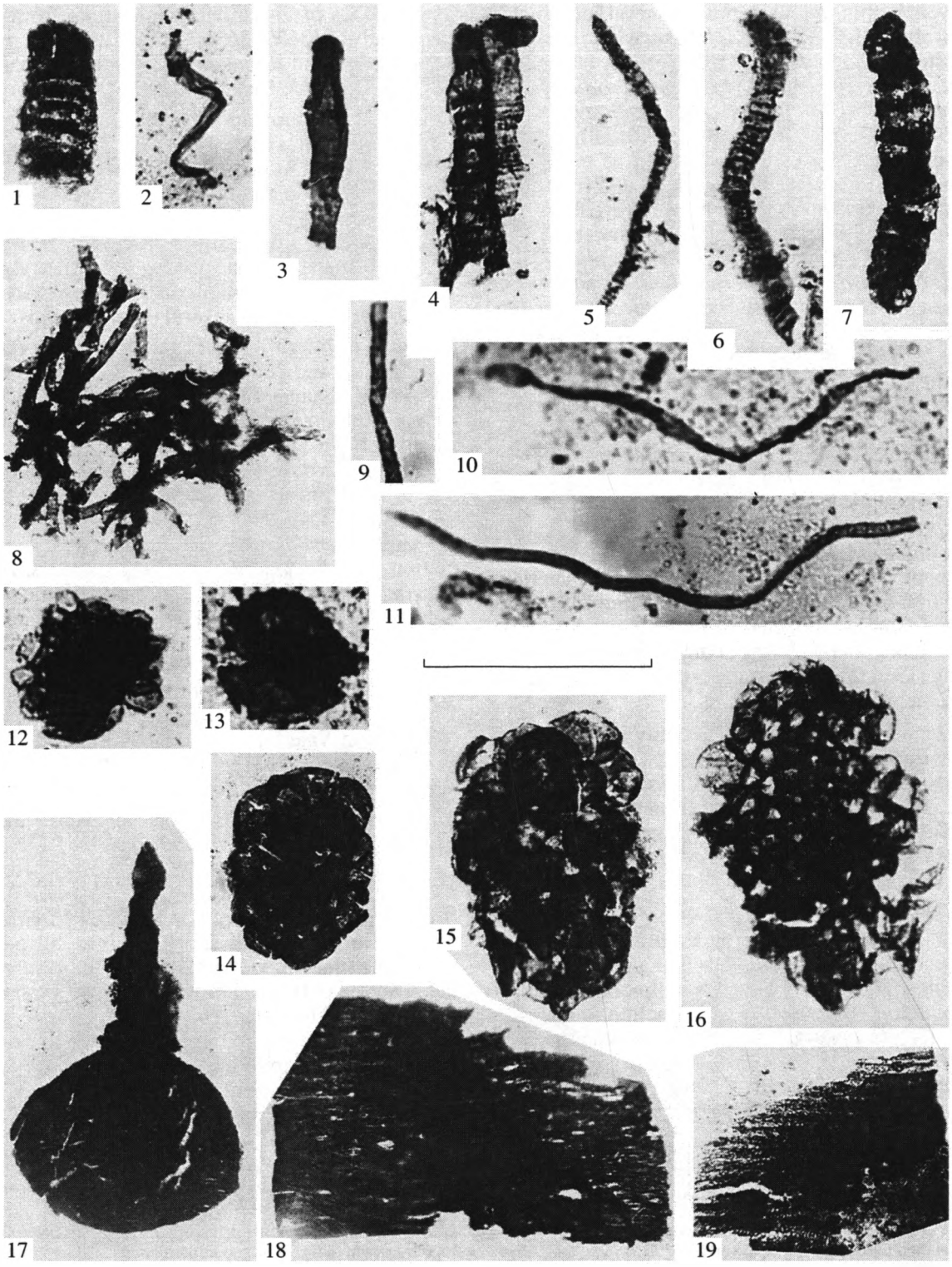
Recent reexaminations and recurrent attempts to find new localities of the Bederysh microbiota insignificantly changed its taxonomic characterization that was elaborated during the 20-year-long period of research. Basic in this case are data on microfossils from three discrete levels of the subformation, which are exposed in four sections spaced from each other, and the fragmentary information about some widespread Riphean morphotypes found in several other outcrops.

Level 6. Microfossils of this level were discovered (Mikhailova and Podkovyrov, 1992) near the village of Bakeevo (Alatau synclinorium, left side of the Zilim River) in the lower shaly beds of the Bederysh Subformation (10 to 30 m above its base). Dominant in the assemblage from this site are medium-sized (100 to 150 μm in diameter) *Leiosphaeridia* species, *L. kulgunica* Jank. and *L. ternate* (Tim.) included, and *Pterospermopsimorpha* forms, both occurring in association with *Arctacellularia*, *Polytrichoides*, *Botuobia*, *Brevitrichoides*, *Leiotrichoides*, and *Tortunema*, which have been formerly known only from the upper subformation part. In addition, we found at this level hercomorphic *Satka* and *Kirbia* formerly unknown from the Bederysh deposits, acanthomorphic *Trachyhystrichosphaera parva* Mikh. and *T. truncata* Herm. et Jank., lophostriate Valeria, very small multiserial coccoidal *Polysphaeroides contextus* Herm., and *Digitus* forms similar to *Archaeoellipsoides*.

Level 7. Microfossils of this level are encountered in the middle shale beds of the subformation (50–70 m below the top) at the Malyi Shishenyak site located 5 km southward of the above locality at the Zilim River (Fig. 1, point 7). Except for single *Trachyhystrichosphaera* fragments with short acanthas, all other acritarchs and filamentous forms of a complex morphology listed above and typical of the lower Bederysh strata are absent here. Their place in microbiota is occupied by taxa of simple morphology, such as size-variable *Leiosphaeridia* (up to 240 μm in diameter), *Myxococcoides* (cells in colonies from 8 to 30 μm in size),

Plate II. Microfossils of the Upper Riphean Karatau Group, Southern Urals (all pictured specimens are stored at the Geological Institute, Moscow)

(1, 2, 7, 8, 12) *Trachyhystrichosphaera aimika* Herm., large envelopes of variable thickness with small tubular processes slightly acute at their endings: (1, 2) specimen no. 9914/501 (magnified envelope fragment in fig. 2), (7) specimen no. 972/133, (8) specimen no. 972/901, (12) specimen no. 972/804; (3, 9, 11) *T. truncata* Herm. et Jank., small and medium-sized compact envelopes with low conical processes: (3) specimen no. 972/176, (9) specimen no. 972/143, (11) specimen no. 972/145; (10) *T. stricta* Herm., medium-sized to large envelopes with tubular processes having filmy acanthas, specimen no. 972/130; (4, 5, 6) *Prolatoforma aculeate* Mikh., ellipsoidal envelopes with bipolar arrangement of acute tubular processes and filmy acanthas: (4) specimen no. 972/123, (5, 6) specimen no. 972/401 (magnified envelope fragment in fig. 5). Specimens by localities: (1,2) Zilmerdak Formation, Bederysh Subformation, microphytological level 8, 3.5 km southward of Min'yar, right bank of the Chernaya River near the Volkovo Village; (3–13) Inzer Formation, lower Subformation, microphytological level 9, Zilim River basin, right bank of the Bol'shoi Shishenyak River 6 km northward of the Kulgunino Village; (2, 12, 15) Zilmerdak Formation, Bederysh Subformation, microphytological level 7, Zilim River basin, left bank of the Malyi Shishenyak River 2 km westward of the Bakeevo Village. Scale bar sizes: (1, 3, 8) 130 μm; (2) 70 μm; (4) 150 μm; (5) 55 μm; (6, 9) 110 μm; (7, 11) 210 μm; (10, 12) 100 μm.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Eomycetopsis and *Leiotrichoides* (2–4 and 8–10 μm wide, respectively). These associate with less abundant other taxa represented by extended *Ostiana* thalli (cells 12 to 14 μm in diameter), small acritarchs *Satka, Nucellosphaeridium, Pterospermopsimorpha* (vesicles not greater than 80 μm in size), and by trichomes *Oscillatoriopsis* and *Cephalophytarion* (up to 10 μm wide).

Level 8. Microfossils from this uppermost stratigraphic level of the Bederysh Subformation represent the well-known Muldakaevo microbiota (Yankauskas, 1982, and later works) that is much more diverse (about thirty genera and species) than those from all lower levels. The microbiota is confined to two members of dark-colored siltstones and shales (8 to 10 m thick each), which are distinguished in sections of the Inzer synclinorium (Iskushta site at the Lemeza River, Fig. 1, point 6) and Adzhigardak anticline (Chernaya Rechka site near Min'yar, Fig. 1, point 5) in the upper part of the subformation.

Composition of microbiota from both sites has been described in a series of publications (Keller and Yankauskas, 1980; Yankauskas, 1982; *Stratotip rifeya...*, 1983; Veis *et al.*, 1990), and, leaving aside the detailed characterization, we list below its key taxa only. These are medium-sized acritarchs *Leiosphaeridia, Pterospermopsimorpha, Nucellosphaeridium*, larger *Chuaria*, catenate coccoidal cells *Chlorogloeopsis (=Enthosphaeroides)*, subspherical or shapeless *Myxococcoides* (*=Synsphaeridium, Microconcentica* and others), sheaths *Eomycetopsis, Leiotrichoides, Polytrichoides, Rectia, Tortunema*, elongated ellipsoidal *Brevitrichoides* with closed endings, shorter akinete-like *Archaeoellipsoides (=Eosynechococcus, Arctacellularia, Lomentunella, Glenobotrydion)* lacking inclusions or enclosing remains of germinating trichomes, abundant and diverse oscillatorian trichomes, beaded *Arctacellularia*, and some other taxa. In the course of our reexamination of two localities (over 20 samples from the Lemeza River section and 40 samples from the Chernaya Rechka site), we identified two additional morphotypes of acanthomorphic acritarchs occurring at the latter site only. These are *Trachyhystrichosphaera aimika* Herm. (110 μm in diameter) with short acanthas slightly broadened near their base and *T. stricta* Herm. (130 × 160 μm in size) with small narrow acanthas rooted in a thin filmy rim. Other newly established morphotypes are large dense opaque thallomes (65 μm wide) and lighter filaments with a longitudinal structure (presumable aggregates of finest sheaths), which resemble *Eomicrocoleus*.

In addition to four aforementioned localities, where the Bederysh Subformation is exposed in its whole range, we collected more than 60 samples within the distribution area of deposits under consideration beginning from the Bol'shoi Nugush River basin in the south to the Berezovskaya anticline in the north of the Bashkirian meganticlinorium. These samples yielded only single specimens of identifiable microfossils, which characterize small outcrops of the Bederysh shales near Inzer (Inzer synclinorium, Fig. 1, point 9) and Bederysh (Berezovskaya anticline, Fig. 1, point 8) settlements and exposures along the Zainysh Creek (a tributary of the Zilim River, Alatau anticlinorium, Fig. 1, point 24). The first and two other localities correspond to microphytological levels 7 and 8, respectively, and yielded identical assemblages of transit Riphean microfossils, which consist of small acritarchs *Leosphaeridia*, coccoidal *Myxococcoides*, and filamentous sheaths *Eomycetopsis* and/or *Leiotrichoides*.

Two assemblages of microfossils, which have been found by Yankauskas in core samples from Borehole Kabakovo-62 (Yankauskas, 1980, 1982; Keller and Yankauskas, 1980; Andreev *et al.*, 1981; *Stratotip rifeya...*, 1982, 1983; *Mikrofosslii...*, 1989) and never revised afterward, presumably characterize the Bederysh interval and are of special interest. They differ from each other in stratigraphic position and taxonomic composition. The lower one (depth interval of 3639–

Plate III. Microfossils of the Upper Riphean Karatau Group, Southern Urals (all pictured specimens are stored at the Geological Institute, Moscow)

(1, 4–7) *Oscillatoriopsis* spp., size-variable trichomes necked near transverse septae with cells of different morphology: (1) specimen no. 972/331, (4) specimen no. 972/410, (5) specimen no. 972/137, (6) specimen no. 972/101, (7) specimen no. 972/101; (2, 3, 8, 9–11) *Leiotrichoides typicus* Herm. emend. Herm., individual specimens and accumulations of simple filamentous sheaths: (2) specimen no. 994/325, (3) specimen no. 972/411, (8) specimen no. 972/174, (9) specimen no. 972/175, (10) specimen no. 997/110, (11) specimen no. 997/111; (12, 14–16) *Myxococcoides* spp., spheroidal and ellipsoidal accumulations of subspheroidal cells: (12) specimen no. 994/316, (14) specimen no. 972/401, (15) specimen no. 994/331, (16) specimen no. 972/114; (13) *Leiosphaeridia bicrura* Jank., small envelope with a characteristic median cleft, specimen no. 973/201; (17) *Caudosphaera expansa* Herm. et Tim., subspheroidal envelopes with long processes (presumable reproductive structures of ancient algae and/or fungi), specimen no. 972/127; (18, 19) *Plicatidium latum* Jank., fragments of cuticular tubular morphotypes (sheaths) with a distinct transverse sculpturing: (18) specimen no. 976/701, (19) specimen no. 972/901. Specimens by localities: (1, 3–9, 16, 17, 19) Inzer Formation, lower Subformation, microphytological level 9, Zilim River basin, right bank of the Bol'shoi Shishenyak River 6 km northward of the Kulgunino Village, (18) Inzer Formation, upper Subformation, lower sequence, microphytological level 10, left bank of the Zilim River 3 km downstream of the Zainysh Creek mouth; (13) Inzer Formation, upper Subformation, upper sequence, microphytological level 11, northern margin of the Novoseitovo Village, mouth area of a nameless tributary of the Saitbak Creek; (14) Min'yar Formation, lower Subformation, microphytological level 12, a quarry near the Byanka railway station; (10, 11) Uk Formation, lower Subformation, microphytological level 16; left bank of the Zilim River 1.6 km downstream of the griffin. Scale bar sizes: (1, 7, 14) 90 μm; (2, 5, 13, 15) 80 μm; (3, 16) 120 μm; (4, 6) 70 μm; (8, 18) 160 μm; (9, 10) 190 μm; (11) 100 μm; (12) 40 μm; (17, 19) 250 μm.

3636 m) is impoverished and resembling to a limited extent, according description, the representative Muldakaevo microbiota known at that time from the Zilmerdak (Bederysh) strata of the Upper Riphean type section (microphytological level 8 in this work). In distinction, the more diverse upper assemblage (depth interval of 3528–3526 m) has been described as including some taxa characteristic not only of the Muldakaevo (upper Bederysh), but also of the Shishenyak (Podinzer) microbiota from a much higher stratigraphic level in the Karatavian sections of southern Urals.

The former assemblage detected near the top of the Priyutovo (Karmaly) Formation (depth interval of 3960–3612 m), which is predominantly composed of variegated siltstones and sandstones, includes large *Leiosphaeridia* (over 200 μm in diameter), oscillatorian trichomes of variable size and morphology, sheaths *Eomycetopsis* and *Leiotrichoides* (2–3 and 10–12 μm wide, respectively), multiserial *Polytrichoides*, compact *Siphonophycus* (up to 40–50 μm wide), ellipsoidal *Brevitrichoides*, and catenate *Chlorogloeopsis* (=*Entosphaeroides*). As was concluded, the assemblage of above taxa can be confidently correlated with that "found in the Bederysh deposits of the Lemeza River basin" (Andreev *et al.*, 1981, p. 60), although at present it can be regarded as an impoverished analogue of any representative Karatavian microbiota.

The latter assemblage is from a middle interval of the overlying Shikhan (Savaleevo) Formation (depth interval of 3612–3507 m) that is predominantly composed of variegated limy rocks. In addition to all taxa of the lower assemblage and abundant transit forms, this assemblage includes very large (over 500 μm in diameter) spongy acritarchs *Tasmanites*, acanthomorphic *Trachyhystrichosphaera aimika* Herm. and *T. trunctata* Herm. et Jank. (=*Baltisphaeridium* sp.), associated *Octaedrixium, Spumosina, Chlorogloeopsis, Polysphaeroides, Tortunema, Glomovertella* (=*Volyniella*), *Rectia, Plicatidium, Arctacellularia, Brevitrichoides*, and abundant trichomes, the wide filamentous *Palaeolyngbya* and structurally complex *Pellicularia* and *Caudosphaera* (=*Fibularix*) included.

Yankauskas and other researchers failed to find many of these taxa, first of all the abundant *Trachyhystrichosphaera, Tasmanites, Octaedrixium, Spumosina, Pellicularia*, and large *Palaeolyngbya*, in the Bederysh deposits and considered them as important components of the younger lower Inzer (Shishenyak) microbiota. It is remarkable therewith that the Shikhan (Savaleevo) assemblage that is most representative was attributed with reservations to the Katav level above the Bederysh Subformation (Andreev *et al.*, 1981). Exactly this scheme of correlation between subdivisions mentioned above was approved recently (*Stratigraficheskaya...*, 2000).

Katav Formation

In the Bashkirian meganticlinorium, the Katav Formation is predominantly composed of variegated and gray limestones promising nothing for microphytological analysis. Yankauskas studied clayey varieties of these rocks in many outcrops, and we sampled the basal, uppermost, and middle intervals of the formation at the Min'yar, Inzer, and Katav-Ivanovsk localities, respectively (Fig. 1, points 10, 11, and 30). All the samples studied are barren of microfossils.

Inzer Formation

The carbonate-terrigenous succession of this formation is of a complex structure. Microfossils are known to occur in inner and outer facies zones at three stratigraphic levels (microphytological levels 10 to 11 in the general succession). According to available data, siliciclastic rocks of the outer zone, where the Inzer Formation is composed of alternating sandy-clayey and carbonate members, yield much more diverse and abundant microfossils than sandstones prevailing in sections of the northern (Lemeza River), central (Inzer Settlement), and southern (Bol'shoi Nugush River basin) areas of the Inzer synclinorium located in the inner facies zone.

It is difficult to interpret unambiguously the vertical succession of Inzer microfossil assemblages because of two reasons. First, they are known from discrete parts of the formation exposed in various tectonic structures, and second, there are controversial views on the relationship between carbonate (lower subformation) and terrigenous-carbonate (upper subformation) strata of the Inzer succession (Komar, 1978; Krylov, 1983; Raaben, 1975, 1981, 1985), which are settled now only to a certain extent (Kozlov, 1986, 1990; Kozlov *et al.*, 1991, 1994, 1997; Maslov and Krupenin, 1991; *Unifitsirovannye...*, 1993; Maslov, 1997, 2002). That is why the succession of Inzer microfossils described below may be subjected to a partial revision in the future.

Lower subformation (Level 9). Yankauskas (1978, 1980) was first to describe the oldest and most representative assemblage of the Inzer microfossils (Podinzer or Shishenyak microbiota) from basal beds of this subformation, which are exposed in the southern part of the Alatau anticlinorium along the Bol'shoi Shishenyak River 6 km northward of the Kulgunino Settlement (Fig. 1, point 12). Later on, the locality was reexamined (Mikhailova and Podkovyrov, 1992). In both periods, abundant microfossils have been macerated from dark-colored shales of the Iken Member (12 to 15 m thick) that overlies rocks of the Katav Formation and grades upward into a complex terrigenous-carbonate succession of the lower Inzer Subformation (*Stratotip rifeya...*, 1982, 1983). According to original description by Yankauskas, the assemblage includes abundant acritarchs *Leiosphaeridia* (*L. bicrura, L. ternate*, and *L. kulgunica* included), *Chuaria, Tasmanites, Leiofusid-*

ium, Simia (=Pterospermella), Pterospermopsimorpha, Nucellosphaeridium, small catenate aggregates of coccoidal *Polysphaeroides* and *Myxococcoides (=Synsphaeridium)*, filamentous sheaths *Eomycetopsis, Heliconema, Polytrichoides, Tortunema*, trichomes *Caudiculophycus, Palaeolyngbya, Calyptothrix, Oscillatoriopsis*, and some other taxa, which are out of use in the current nomenclature. Later on, Mikhailova added to this list such taxa as *Satka, Germinosphaera, Leiotrichoides, Archaeotrichion, Botuobia*, and *Clavitrichoides*, all commonly occurring in Riphean deposits. In addition, she was first to identify exactly at this level of the Karatavian type section the abundant acanthomorphic *Trachyhystrichosphaera* species (*T. parva* Mikh., *T. aimika* Herm., *T. truncata* Herm. et Jank., and *T. vidalii* Knoll), described the new genus *Prolatoforma* (with a single species *P. aculeate* Mikh.), and distinguished the peculiar gemmate *Eosaccharomyces*, the presumable remains of yeast-like fungi (Mikhailova and Podkovyrov, 1992).

We studied the Bol'shoi Shishenyak locality in detail (over 20 samples with abundant microfossils have been analyzed). According to our results, the Iken Member of siltstones and shales that is comparatively thin here includes three regularly spaced intervals, within which microfossils are similarly abundant and diverse. Within the in-between intervals (4–6 and 10–12 m above the member base), as well as near the base and top of the member, microfossils are much less abundant and diverse.

In addition to taxa reported by our predecessors, we identified among the Shishenyak acritarchs a series of new morphotypes confined to all three levels mentioned above. These are large ellipsoidal *Konderia* (140 × 260 μm), compact spongy *Spumosina* (vesicles 70 to 170 μm in size), *Aimia* forms (up to 50 μm in size) with several internal bodies, sheaths similar to *Cerebrosphaera* (up to 200 μm in diameter) with a convolution-type surface, acanthomorphic *Trachyhystrichosphaera stricta* Herm. (from 140 to 230 μm in diameter) with a narrow filmy rim enveloping small acanthas, *Cymatiosphaeroides* (up to 180 μm in size) with numerous thin processes, sheaths of irregular shape (90 to 130 μm) presumably representing monstrous morphotypes (Veis *et al.*, 1999), and subsphaerical accumulations of Leiosphaeridia (20, 30, and 50–60 μm in size). Among colonial coccoidal forms previously unknown at the 9th level, we recognized shapeless *Ostiana* thallomes (cell diameters 10–12 and 18–20 μm), multiserial catenate aggregates *Chlorogloeopsis* (cells 12 to 14 μm in size), and botryoidal cell clusters resembling *Palaeopleurocapsa* (cell diameter 20 μm).

Besides *Eomycetopsis* and *Leiotrichoides* known before, the diverse filamentous forms of the Shishenyak microbiota are represented by fragments of their compact mats, by wider unornamented *Asperatofilum* (up to 40 μm) and *Taenitrichoides* (90–100 μm), by costate *Rectia* (30–40 μm), by two generations of very large (140 and 260–280 μm) *Plicatidium* with a fine transverse striation, and by elongated aggregates of thin sheaths *Eomicrocoleus* (some as wide as 50 μm). The trichome-like morphotypes formerly unknown in the Shishenyak microbiota are catenate *Arctacellularia* consisting of nearly isometric cells (40 × 45 μm), *Trachytrichoides* with elongated cells (18–20 × 24–26 μm), and unusual elongated forms composed of large cells-segments (100 μm wide and 80, 100, and 120 μm long). Microfossils of a complex structure from microphytological level 9 are small fusiform *Fabiformis* (160 × 280 μm), vesicles *Caudosphaera* with a long solitary process (250 μm in diameter, process 200 μm long), and compact opaque thallomes of ribbon-like shape (40, 70, 100, and 120 μm wide). Thus, our research increased the Shishenyak microbiota diversity from 22 to 34 genera (more than by one third) and from 28 to 52 species (by two times approximately).

According to published data (Keller and Yankauskas, 1980, and other works), the less diverse microbiota from a southern locality (exposures of the "Shishenyak Member" along the Uryuk River, Yalmash-Uryuk anticline) is similar in general to the microbiota from basal Inzer strata of the Alatau anticlinorium, which is described above. However, we exclude it from consideration in this work, because illustrations of taxa and the exact data on stratigraphic position and taxonomic composition of that microbiota have not been reported in the mentioned publication.

Microbiota of the level 9 probably includes, in addition to forms characterizing the representative locality Bol'shoi Shishenyak, also the less abundant taxa known from the Karatau structural complex (Karatau Ridge anticline) at the Bannyi Creek locality (Fig. 1, point 31) near the village of Ivanovka (Veis *et al.*, 1990). Taxa found here in the upper part of the Iken Member (10–12 m below its top) are small *Leiospaeridia* (vesicles up to 60 μm in diameter), extended *Myxococcoides* colonies of unusual palmelloid appearance formerly attributed to *Eoentophysalis* (cells 9 to 10 μm in diameter), thin and narrow filamentous sheaths *Leiotrichoides* (8 to 10 μm wide), wider compact *Siphonophycus* (24 to 26 μm), trichomes *Oscillatoriopsis* (12 μm wide), and fusiform *Nelcanica*, which consist of several embedded sections thinning at their lighter rounded endings (30–40 × 100–110 μm in size).

Upper subformation, lower member (Level 10). Five microfossil assemblages characterize this level, three of which are impoverished, slightly different in taxonomic composition. The latter are established in the central part of the Alatau anticlinorium in exposures of the Podinzer Beds along the Zilim River. We found two assemblages of this kind downstream of the Zainysh Creek mouth (Fig. 1, point 13), and one was found by Mikhailova (Mikhailova and Podkovyrov, 1992) near the Bala-Elga Creek mouth (Fig. 1, point 32), where it is in a direct contact with microbiota of the next microphytological level. All three assemblages are

from beds, which are separated from the Katav limestones with characteristic microbial lamination by several lower Inzer (upper Katav after Krylov, 1983) beds (each a few tens of meters thick) of "algal-laminated" and clastic limestones. At the Zainysh Creek localities, microfossils are confined to a package (20 m thick) of fine-grained sandstones intercalated with gray to variegated siltstones and shales. Gray shales yield here abundant acritarchs *Leiosphaeridia* (30 to 280 μm in diameter), less abundant *Chuaria* (380–400 μm), *Nucellosphaeridium* (up to 140 μm), and oval *Navifusa* (120–140 × 300–320 μm), some of which are with a longitudinal striation similar to that of *Cucumiforma*. These morphotypes occur in association with the following taxa: large vesicles with several included bodies *Aimia* (320 to 340 μm in diameter), colonial coccoidal *Myxococcoides* (cells up to 20 μm in diameter), filamentous sheaths *Eomycetopsis*, aggregated *Eomicrocoleus* (2–3 and 20–24 μm wide), and very wide (up to 280 μm) fragmented *Plicatidium* having a fine transverse sculpturing. According to data of Mikhailova, forms dominant at the Bala-Elga Creek locality are medium-sized *Leiosphaeridia*, *Pterospermopsimorpha*, peculiar flaked *Kirbia*, acanthomorphic *Trachyhystrichosphaera truncata* Herm. et Yank., and fragments of oscillatorian trichomes.

Besides the Zilim River sites, there are two others localities in limbs of the Berezovskaya (Sim River) and Adzhigardak (town of Min'yar) anticlines, which yield microfossils of the level 10. In both structures remote from each other, microfossils occur at the top of the lower sandy-clayey member of the upper Inzer Subformation below the middle Minka (*Stratotip rifeya...*, 1983; Raaben, 1985, 2001) or Priprudnaya (Krylov, 1983) Member of characteristic carbonate rocks with abundant stromatolites *Gymnosolen ramsay* Steinm. In the former structure (Sim River site downstream of the Solenyi Creek, Fig. 1, point 34), the assemblage found in two closely spaced intervals of the section (15 to 20 m thick each) consists of abundant acritarchs *Leiominuscula*, *Leiosphaeridia*, and *Nucellosphaeridium* (8 to 70 μm in diameter), small coccoidal *Myxococcoides* (cells 6 to 7 μm in size), and of filamentous sheaths *Eomycetopsis* and *Leiotrichoides* (from 2 to 16 μm wide).

In the latter structure, microfossils are found in a small, less than 10 m thick outcrop of shales (lower strata of the upper subformation or the Telegorsk Beds after Krylov, 1983) situated near the railway between the Min'yar and Byanka railway stations (Fig. 1, point 33). In addition to many taxa known from the upper Inzer Subformation at the Sim River and Min'yar localities, there were found less abundant but larger acritarchs *Leiosphaeridia* (up to 160 μm in diameter) and *Chuaria* (up to 400 μm in diameter) in association with sheaths *Siphonophycus* (26 to 28 μm wide).

Upper member (level 11). This level is distinguished at five localities. One of them corresponds to thin packages (a few meters thick) of gray siltstones and shales of the upper Inzer Subformation, which are exposed near the village of Novoseitovo (Fig. 1, point 14) in the southern part of the Inzer synclinorium. The packages are situated 70 to 80 m above the top of the lower Inzer Subformation (*Stratotip rifeya...*, 1983; Maslov, 1997) and yields single *Leiosphaeridia* forms (*L. bicrura* and *L. ternata*), *Nucellosphaeridium* (from 20 to 40 μm in size), *Myxococcoides* (cells 8 to10 μm in diameter), and *Eomycetopsis* (3 to 4 μm wide).

Two other localities are found in northern (small quarry southwestward of the Lemeza Settlement, Fig. 1, point 27) and central (a long road cut in the western Inzer suburb near the Ufa-Beloretsk highway, Fig. 1, point 26) parts of the Inzer synclinorium (Veis *et al.*, 1990). At both sites, microfossil assemblages of low diversity are confined to greenish gray shale interlayers (5 to 10 cm thick) in fine-grained sandstones prevailing here. The interlayers yield only small *Leiosphaeridia* (60 to 75 μm in diameter) and/or *Leiotrichoides* (8 to 10 μm wide).

One more locality was found near the village of Bakeevo (Fig. 1, point 23) close to the Inzer Formation top (Mikhailova and Podkovyrov, 1992). Microfossil taxa identified here are small *Leiosphaeridia*, *Pterospermopsimorpha*, *Kirbia*, *Trachyhystrichosphaera truncata* Herm. et Jank., *T. parva* Mikh., compact spongy *Spumosina*, and single oscillatorian trichomes.

The last assemblage of the level 11 is likely terminal one in the succession of upper Inzer microbiotas established at present. It is found near the subformation top in shale beds (about 10 m thick) immediately below basal dolostone beds of the Min'yar Formation (Adzhigardak anticline, Byanka quarry near Min'yar, Fig. 1, point 15). The assemblage includes acritarchs *Leiosphaeridia* (from 20 to 260 μm in diameter), coccoidal *Myxococcoides* (cells 8–10 and 20–22 μm across), sheaths *Eomycetopsis* (3 to 4 μm wide), and fragments of spongy fusiform *Fabiformis* (120 × 350 μm). A similar assemblage of taxa was discovered here by Yankauskas (Keller and Yankauskas, 1980; *Stratotip rifeya...*, 1982).

Min'yar Formation

Lower subformation. Predominant dolostone beds of the subformation are intercalated with rare and thin (a few millimeters to dozens of centimeters thick) interlayers of shales, siltstones, and marls, which may bear the organic-walled microfossils. The latter are known from two stratigraphic levels of the Min'yar Formation (microphytological levels 12 and 13 in the general succession). **Level 12** is established near the base of the Adzhigardak Subformation that is exposed well in the Byanka quarry (Fig. 1, point 15) near town of Min'yar (Veis *et al.*, 1990). Some interlayers of dark colored marls (5 to 10 cm thick) yield here rare acritarchs *Leiosphaeridia* (*L. kulgunica* Jank. included) up to

50 μm in diameter, single coccoidal *Myxococcoides* (large cells 20 to 34 μm across), and filamentous sheaths *Leiotrichoides* (up to 10 μm wide).

Microfossils of **Level 13** are found in black shale interlayers a few millimeters thick, which are exposed 10 to 15 m above the previous level in the same quarry (Fig. 1, point 15). Like the previous one, this level yields a very impoverished assemblage of microfossils. The latter includes acritarchs *Leiosphaeridia* of the same size range, *Nucellosphaeridium* (50 and 150 μm in diameter), *Chuaria* (320–340 μm across), and filamentous sheaths *Leiotrichoides* (8 to 10 μm wide).

Upper subformation. Organic-walled microfossils are unknown from this subformation lacking distinct shale interlayers. Nevertheless, the well-preserved and abundant mineralized microfossils have been found in dark nodules and lentils of early diagenetic cherts widespread in upper beds of stratified and clastic dolostones corresponding to the Balaevo and Kama-Elga members (Krylov, 1983; Nyberg and Schopf, 1984; Sergeev and Krylov, 1986; Sergeev, 1992). They are confined to a narrow stratigraphic intervals in four localities situated at the Inzer, Zilim, and Kuryak rivers and near the town of Min'yar. Two assemblages of different composition from this interval characterize the microphytological **Level 14.** Their taxa are coccoidal *Gloeodiniopsis* (up to 40 μm), *Eoaphanocapsa* (9 to 17 μm), *Eoentophysalis* (2 to 9 μm), *Sphaerophycus* (3 to 6.5 μm), *Eogloeocapsa* (18–15 μm), acritarchs *Leiosphaeridia* (30 μm), small ellipsoidal *Eosynechococcus* (3 × 6 μm), empty sheaths *Eomycetopsis* (2 to 4.5 μm), *Siphonophycus* (4.5 to 33 μm), sheath *Entosphaeroides* with internal bodies (3–4 μm), branching filamentous *Ramivaginalis* (4–9 μm), dark filamentous *Biocatenoides* (0.5 to 1 μm), *Palaeolyngbya* (23–30 μm), and trichomes *Oscillatoriopsis* and *Caudiculophycus* (12–14 and 1.5–3 μm, respectively). We also arbitrarily attribute to this level rare specimens of vase-shaped *Melanocyrillium*, which have been reported from an upper part of the Min'yar Formation near a pond in the town of Yuryuzan (Maslov *et al.*, 1994) but not confirmed so far anywhere else.

Uk Formation

Lower subformation. Our predecessors and we sampled this subdivision in many sections of the Karatau Group situated in western and eastern limbs of the Bashkirian meganticlinorium. All the data available at present show that microfossils attributable to microphytological levels 15 and 16 occur only in sandstone-siltstone beds of the lower Uk Subformation exposed along the Zilim River within a central area of the Alatau anticlinorium.

Microfossils of the **Level 15** correspond to those found by Yankauskas in a small outcrop near the village of Bakeevo at the Zilim River (Fig. 1, point 28), where a basal portion of the lower Uk Subformation is exposed. This locality is one of the most representative among other outcrops of the Uk Formation known so far. It yields the following taxa: abundant acritarchs *Leiosphaeridia* and *Nucellosphaeridium* of a medium size; small coccoidal *Polysphaeroides* and *Myxococcoides*; smooth-walled and ornamented filamentous sheaths *Eomycetopsis, Leiotrichoides, Polytrichoides, Tortunema, Siphonophycus*, and *Rectia*; rare oscillatorian trichomes; trichome-like *Arctacellularia*; and fusiform *Pellicularia* of a complex structure (Keller and Yankauskas, 1980; *Stratotip rifeya...*, 1982).

Level 16 is separated from the previous one by a poorly exposed section interval a few tens of meters thick. At the Zilim River, we found two microfossil assemblages of the level, which are conjugate in lateral direction, in dark colored siltstones and shales of the lower Uk Subformation. One assemblage from a small outcrop situated 1 km downstream from the Zilim Griffin (Fig. 1, point 17) consists of abundant acritarchs *Leiosphaeridia* (30 to 240 μm in diameter), small *Chuaria* (300–320 μm), *Nucellosphaeridium* (up to 200 μm across), small *?Tetrasphaera* (up to 30 μm), fragments of *Trachyhystrichosphaera parva* Mikh. (up to 45 μm across), and compact *Leiosphaeridia* accumulations (some vesicles are 50–80 μm in diameter). Colonial coccoidal morphotypes are represented by small thallomes *Ostiana* (cells 22–24 μm in diameter) and by *Myxococcoides* of two size generations (16–18 × 20–22 μm). Filamentous sheaths represent either individual specimens of *Eomycetopsis* (3 to 4 μm wide) and *Leiotrichoides* (8 to 10 μm wide), or mats of the former taxon. There are also very large trichome-like morphotypes (up to 80 μm wide) in this assemblage.

The second assemblage from a similar stratigraphic position is from a local outcrop of the lower Uk shales and siltstones, which is located at the Zilim River 2 km downstream of the Zainysh Creek mouth (Fig. 1, point 29). This outcrop yields only small acritarchs *Leiominuscula, Leiosphaeridia, Pterospermopsimorpha* (vesicles from 8 to 80 μm in size), coccoidal *Myxococcoides* (cells in colonies 22–24 μm across), and *Leiotrichoides* (6–8 μm wide).

Yankauskas also studied deposits of the Krivaya Luka Formation (its lower sandstone and siltstone beds of the Magadeevo Subformation) that is terminal one in the Upper Riphean succession of the Bashkirian meganticlinorium eastern limb. We also sampled these deposits in the Belaya River valley near the Krivaya Luka ravine and Muradymovo Settlement. The sampled rocks turned out to be highly altered and barren of identifiable microfossils.

The Tolpar and Suir formations,[1] the possible analogues of the Krivaya Luka Formation in a local north-

[1] In opinion of Kozlov *et al.* (1991, 1994, 1997), both formations belong to the Lower Vendian, whereas Mikhailova and Podkovyrov (1987, 1992) attributed the Tolpar and Suir formations to the Upper Riphean and Lower Vendian, respectively. Other researchers (Keller *et al.*, 1984; Veis *et al.*, 1990) conventionally regarded them as subdivisions of the Karatavian. Taking into consideration these controversial interpretations, we do not include the Tolpar and Suir microfossil assemblages into the general succession of Karatavian microphytological levels.

ern depression of the Zilim syncline (the Tolparovo area), represent the other case. Peculiar microfossil assemblages differing from each other in abundance and diversity of their taxa are known from a series of exposures of both formations in the Zilim and Malyi Tolpar valleys. The Tolpar microbiota resembles to some extent the other Karatavian microbiotas of southern Urals, whereas the Suir assemblage is closer in composition to Vendian microbiotas of the Russian plate (*Stratotip rifeya...*, 1983; Keller *et al.*, 1984; Veis *et al.*, 1990; Mikhailova and Podkovyrov, 1987, 1992). Not frequent siltstone interlayers in sandstones of the Tolpar Formation yield abundant acritarchs *Leiosphaeridia* (*L. kulgunica* Jank. included), small *Valeria, Satka, Pterospermopsimorpha, Trachyhystrichosphaera parva* Mikh., *Spumosina*, and less abundant coccoidal *Myxococcoides* and *?Sphaerocongregus*, all associated with filamentous sheaths *Eomycetopsis, Leiotrichoides*, and *Polytrichoides*. Taxa known from the Suir Formation composed of thin sedimentary rhythms of flyschoid clayey rocks are *Leiosphaeridia, Spumosina*, very small *Bavlinella* and *Retiforma*, coccoidal *Myxococcoides*, and thin filamentous sheaths *Pomoria* with characteristic "rhombic" decoration.

MAIN CONCLUSIONS

Elaborating the original well-known scheme of microphytological subdivision of the Riphean type section, Keller and Yankauskas (1980) took into consideration all the data on microfossils from Riphean deposits recovered by boreholes in eastern areas of the Russian plate (Kabakovo-62 and others). Correlation of these deposits with the Riphean stratotype was based on lithostratigraphic, paleontological, and geophysical materials interpreted sometimes in a controversial manner (Andreev *et al.*, 1981). Afterward, in addition to the scheme adopted for the southern Urals proper (*Stratotip rifeya...*, 1982), there was elaborated the general stratigraphic scheme of the Upper Precambrian (Yankauskas, 1982) that integrated the results of microfossil study in southern Urals, Bashkirian and other Cis-Urals areas was authorized as leading one for biostratigraphic research here (*Mikrofossilii...*, 1989; *Unifitsirovannye...*, 1993; *Stratigraficheskaya...*, 2000). In this general scheme, the Riphean includes six discrete and nonuniform intervals of the section ranging, in terms of stratigraphic nomenclature, from a group to a subformation and representing in essence the auxiliary biostratigraphic subdivisions, i.e. the horizons with microfossil assemblages, which are unequally diverse and meaningful. These are the Burzyan Horizon of the Lower Riphean, the Yurmata Horizon of the Middle Riphean, the Kabakovo, Muldakaevo, and Shishenyak horizons of the Upper Riphean; and the Kudash Horizon attributed to the terminal Riphean.

Analyzing the succession of microfossils in the Upper Riphean deposits of southern Urals, Yankauskas (see in *Stratotip rifeya...*, 1982) concluded that distinctions between the upper Zilmerdak (Muldakaevo) and lower Inzer (Shishenyak) microbiotas are significant enough to recognize two lower subdivisions of the Karatau Group, which can be regarded as comparable with the horizons used in practice of Paleozoic stratigraphy. Distinctive taxa of the former microbiota were assumed to be filamentous *Arctacellularia, Cephalophytarion, Polytrichoides, Tortunema*, and *Glenobotrydion (=Archaeoellipsoides)* occurring in association with less abundant acritarchs *Leiosphaeridia kulgunica* Jank. and, sometimes, with *Chuaria* and *Tasmanites*. Taxa regarded as dominant in the latter microbiota are acritarchs *Chuaria, Tasmanites, Pterospermella (=Simia)*, and *Leiofusidium* accompanied by subordinate sheaths and trichomes *Eomycetopsis, Heliconema, Oscillatoriopsis, Caudiculophycus*, and *Palaeolyngbya*.

In the 1970s–1980s, microfossils of the Middle Karatavian interval (Inzer and Min'yar formation) were inadequately known, and it was impossible to distinguish here biostratigraphic subdivisions. According to data of that time, the youngest Uk microbiota from the Upper Karatavian interval was less diverse than the Muldakaevo and Shishenyak microbiotas. It was reported to include, in addition to a series of transit taxa, the abundant simple sheaths *Eomycetopsis* and peculiar "rounded" trichomes *Oscillatoriopsis (=Palaeolyngbya) zilimica* (Jank.) appearing at that level in association with wide tubular *Rectia (=Siphonophycus)* having a transverse ornamentation. In opinion of Yankauskas (1982), these distinctions are insufficient to suggest a considerable microphytological renovation before appearance of the Uk microbiota.

Nevertheless, in the general scheme of the Karatau Group biostratigraphy (Yankauskas, 1982), the above microbiotas representative to a different extent were regarded as characterizing the Muldakaevo, Shishenyak, and Kudash microphytological horizons separated from each other by barren intervals. The Kabakovo Horizon distinguished in synonymous borehole section was placed at the base of the composite Upper Riphean succession. Microfossils of the last horizon were those from the Tukaevo and Ol'khovyi (lower part) formations (in modern terminology), and very large lophostriated acritarchs *Valeria* and some chercomorphic *Satka* represented distinctive taxa of their assemblage. As was suggested by Yankauskas (1982), the Kabakovo level corresponds in the Upper Riphean type section to the lowermost Biryan Subformation of the Zilmerdak Formation, which is lacking meaningful paleontological remains.

In accord with the declared approach to microphytological subdivision of the Riphean, it was assumed that each of the above horizons is preceded by a corresponding event of microbiota renovation of an interregional significance. In practice however, a significant renovation was demonstrated for the pre-Muldakaevo event only, especially when it was decided based on

correlation that the Muldakaevo microbiota includes *Trachyhystrichosphaera, Plicatidium, Chlorogloeopsis, Octaedrixium*, and other important morphotypes found in Borehole Kabakovo-62 (depth interval 3528–3526 m) and characteristic of the Lakhanda, Miroedikha, and some other microbiotas of Siberia. The pre-Shishenyak event in southern Urals was outlined conventionally by analogy with Siberian (Ust-Kirba microbiota) and Scandinavian sections (Yankauskas, 1982), and the pre-Kabakovo and pre-Uk (pre-Kudash) events have never been traced beyond the Cis-Ural region and southern Urals (*Mikrofossilii...*, 1989). Later on, different views on main events in the Precambrian microbiota history of and controversial interregional correlations based on microfossils, which contradicted sometimes to indications of other methods, inflated significance of the Muldakaevo Horizon. Simultaneously, all other Upper Riphean horizons (except for the Kabakovo Horizon) and, especially, the Burzyan and Yurmata horizons distinguished formally lost their correlation significance, being regarded as levels of discrete microbiotas or of their voluntary combinations (Maslov *et al.*, 2001).

This situation adequately reflects the microphytological specifics of Lower-Middle Riphean deposits in the stratotype area, where they yield predominantly simple transit morphotypes (Veis *et al.*, 1990; Sergeev, 1992, 1994; Sergeev and Lee Seong-Joo, in press), although the real distribution patterns of microfossils in younger Upper Riphean deposits are of another kind. The microphytological characterization of the Karatau group was radically clarified in recent years (Mikhailova and Podkovyrov, 1992; Kozlov *et al.*, 1994; Maslov *et al.*, 1994; Veis *et al.*, 2000; and this work) and opened a possibility to assess objectively the biostratigraphic potential of microfossils in subdivision of the Upper Riphean.

Basic for this assessment is biostratigraphic analysis of ten successive microbiotas known now from the Karatau Group. Four of them (Biryan, Nugush, Lemeza, and Bederysh microbiotas) belong to the Zilmerdak Formation, three to the lower and upper Inzer subformations (one and two, respectively). Two microbiotas characterize the Min'yar Formation (its lower and upper subformations), and one is known from the lower part of the Uk Formation. The absolute majority of morphotypes are represented in these microbiotas by organic-walled microfossils. The latter are represented by 65 genera and 85 species (Fig. 2). Mineralized microfossils are known from the upper part of the Min'yar Formation only. All these forms from 32 localities, nine of which have been discovered in the course of our study, characterize 16 microphytological levels.

In addition to microfossils whose connections with the Karatavian of southern Urals are out of doubts, we considered the Kuzha and Tyul'ma organic-walled microbiotas from the Middle or Upper Riphean deposits of the stratotype area (18 genera and 17 species of the former and 15 genera and 19 species of the latter from seven microphytological levels) and the Tolpar and Suir microbiotas (15 genera and 20 species from two microphytological levels) attributable to the Upper Riphean and/or Vendian. To complete the general distribution of microfossils within the time span under consideration, we also paid attention to the Tukaevo, Ol'khovyi, Priyutovo, and Shikhan microbiotas from synonymous formations of the Serafimovka and Abdulino groups of the Volga-Ural region (in total, 26 genera and 32 species of four microphytological levels). It is necessary to argument additionally the formerly established positions of all nine microbiotas listed above in the Riphean succession of microfossils.

In principle, such an argumentation means an objective consideration of microfossil distribution in the Burzyan, Yurmata, and overlying Karatau groups of the Riphean stratotype and a correct correlation between these and other microbiotas of a doubtful stratigraphic position. This approach means *a priori* a certain autonomy of biostratigraphic conclusions with respect to age ranges of subdivisions, which bear the microbiotas, as compared to those accepted now in regional stratigraphic schemes of the Riphean in the Volga-Ural region (*Stratigraficheskaya...*, 2000) and southern Urals (*Unifitsirovannye...*, 1993). The autonomy degree is assessed quite variably, however, by different researchers (we are certainly among them).

According to their distribution patterns in the Riphean type section, microfossils from the Karatau Group proper are divisible in two groupings, one of which includes about a third of Karatavian taxa (21 genera and 37 species). These taxa represent most simple transit morphotypes, are dominant everywhere, and ascend up to the Karatau Group from underlying Burzyan and/or Yurmata groups (Fig. 2). The grouping consists of small and medium-sized acritarchs, colonial coccoidal, and filamentous microfossils. These are *Leiominuscula, Leiosphaeridia, Nucellosphaeridium, Pterospermopsimorpha, Satka, Myxococcoides*, marginate forms, *Coniunctiophycus, Gloeodiniopsis, Palaeopleurocapsa, Eogloeocapsa, Sphaerophycus, Eomycetopsis, Leiotrichoides, Siphonophycus, Eomicrocoleus, Eosynechococcus, Archaeoellipsoides, Oscillatoriopsis* and other oscillatorian trichomes, *Palaeolyngbya*, and *Enthosphaeroides*. Morphotypes representing the grouping are relatively small: maximal sizes are less than 200 μm for *Leiosphaeridia*, less than 60 μm for *Eogloeocapsa*, less than 200 μm for sphaeronostocaleans, less than 32 μm for *Siphonophycus*, less than 25 μm for *Palaeolyngbya*, less than 15 μm for oscillatorian trichomes, and less than 35 μm for marginate envelopes (*Stratotip rifeya...*, 1982; *Mikrofossilii...*, 1989; Veis *et al.*, 1990; Mikhailova and Podkovyrov, 1992; Sergeev, 1992, 1994; Sergeev and Lee Seong-Joo, 2003).

Large morphotypes and/or those of a complex structure, which are unknown outside the Karatau Group in the Riphean type section, belong to the second more diverse grouping that includes all principal varieties of Riphean microfossils (44 genera and 53 species). These are *Chuaria, Tasmanites, Tetrasphaera, Konderia, Simia, Trachyhystrichosphaera, Cymatiosphaeroides, Prolatoforma, Kirbia, Aimia, ?Cerebrosphaera, Cucumiforma*, monstrous (fissionable) envelopes, *Polysphaeroides, Plicatidium, Rectia, Tortunema, Glomovertella, Heliconema, Arctacellularia, Botuobia, ?Proterocladus, Eosaccharomyces, Majaphyton, Fabiformis, Nelcanica, Caudosphaera, Lakhandinia, Pellicularia, Melanocyrillium*, giant trichome-like and some other morphotypes. Maximal sizes of microfossils from this grouping are 1000 μm for acritarchs, up to 300 μm for nonseptate filamentous forms, 100 μm for trichome-like morphotypes, and 500 μm for branching thallomes.

In addition, the Karatavian and some other, presumably Late Riphean microbiotas (those from the Kuzha, Tyul'ma, Tolpar, and Suir formations) include other taxa, which are similarly large. These are *Valeria, Navifusa, Ostiana, Sphaerocongregus, Asperatofilum, Taenitrichoides, Polytrichoides, Octaedrixium, Spumosina, Navifusa, Germinosphaera, Brevitrichoides, Pseudodendron, Trachytrichoides*, very wide *Siphonophycus*, and well identifiable *Palaeolyngbya sphaerocephala* Herm. et Pyl. (Fig. 2).

Corymbococcus from the lower Satka level and *Biocatenoides, Ramivaginalis*, and some other morphotypes from the upper Min'yar Subformation are unique to a certain extent. In the study region, they are unknown beyond the indicated subdivisions but have been described from various Riphean and pre-Riphean strata in other regions (Sergeev, 1992; Sergeev and Lee Seong-Joo, 2003).

Stratigraphic and geographic distribution of morphotypes representing the second groupings are irregular in the Karatavian type area. The most diverse microbiotas are known here from the Nugush, Bederysh, and lower Inzer subformations (Fig. 2), which yield near the base the taxa unknown below in the section but rising up into overlying subdivisions. In the Nugush Subformation, newcomers of this kind are *Chuaria, Navifusa, Simia nerjenica* A. Weiss, monstrous (fissionable) sheaths, *Ostiana, Polysphaeroides contextus* Herm., *Asperatofilum*, wide ornamented *Siphonophycus, Taenitrichoides, Plicatidium, Rectia, Botuobia*, giant trichome-like morphotypes, *Majaphyton, Lakhandinia, Caudosphaera*, presumable *Proterocladus* and *Pseudodendron*. Taxa newly appearing in the Bederysh Subformation are *Leiosphaeridia kulgunica* Jank., *Valeria, Trachyhystrichosphaera* (*T. parva* Mikh., *T. trunctata* Herm. et Jank., *T. aimika* Herm., *T. stricta* Herm.), *Chlorogloeopsis, Brevitrichoides, Tortunema, Glomovertella, Polytrichoides*, and *Arctacellularia*. In the lower Inzer Subformation, the first occurrence is recorded for giant *Chuaria, Tasmanites, Konderia, Spumosina, Simia simica* (Jank.), *Trachyhystrichosphaera vidalii* Knoll, *Prolatoforma, Cymatiosphaeroides, Kirbia, Aimia, Cerebrosphaera, Cucumiforma, Heliconema, Trachytrichoides, Eosaccharomyces, Fabiformis*, and *Nelcanica*.[2] In the lower Uk strata higher in the section, only *Tetrasphaera* known nowhere else and peculiar *Oscillatoriopsis zilimica* (Jank.) can be tentatively attributed to taxa of the same kind. As for the Karatau Group intervals, which separate strata bearing taxa listed above, they yield only impoverished sets of taxa from the first and/or second groupings.

Thus, the microphytological characterization of the Karatavian type section that is known at present confirms the earlier data on distribution patterns of microfossils in the middle section interval. As before, we can distinguish here the pre-Bederysh and pre-Inzer regional events, which predetermined appearance of many taxa unknown in underlying deposits. Situation with microfossil distribution in the upper part of the Karatau Group is somewhat different. The upper Inzer and Min'yar deposits demonstrate a considerable decline in diversity of microbiotas that is restored to some extent only in the lower Uk strata owing to the recurrent penetration of certain taxa from underlying deposits and simultaneous appearance of few newcomers. These taxonomic changes are insufficient to admit objectively the stratigraphic significance of indicated biotic events and do not confirm the idea of Yankauskas (1982) who suggested the pre-Uk compositional transformation of South Uralian microbiotas. The same can be said with respect to impoverished microbiotas from the overlying Tolpar and Suir formations. The first of them includes *L. kulgunica* Jank., *Valeria, Spumosina, Trachyhystrichosphaera, Sphaerocongregus*, and *Polytrichoides* characteristic of the Karatavian strata, whereas the second one is lacking these morphotypes, except for *Spumosina*, and includes *Bavlinella, Retiforma*, and *Pomoria* unknown below in the Riphean succession. In terms of microphytology, it seems reasonable to include the Tolpar assemblage into the succession of Riphean microfossils and to exclude from the latter the Suir microbiota likely belonging to the Vendian (Mikhailova and Podkovyrov, 1992).

Most significant biotic events are recorded in the Karatau Group below the Bederysh Subformation. The Nugush microbiota that appeared here and whose taxa rise up into overlying Karatavian deposits clearly points to the important pre-Nugush level of transformation of South Uralian microbiotas, a possible analogue

[2] According to available sedimentological reconstructions (Maslov, 1997), the appearance levels of the above three newcomers' assemblages correspond to distinct stages of the Karatavian basin deepening and emergence of settings with quiet hydrodynamics. In the Riphean, settings of this type were favorable for development of facies-ecologic grouping of microorganisms, which were especially representative, consisting of most stratigraphically important morphotypes (Veis and Petrov, 1994; Veis *et al.*, 1998a).

of which (pre-Kabakovo level) was earlier recognized in the Cis-Ural region and correlated with the Karatavian base without a valid microphytological substantiation. A stricter localization of the pre-Nugush level (or of levels) in the Riphean succession of the Bashkirian meganticlinorium is impossible, because nearly a half of taxa typical of the Nugush and other Karatavian microbiotas (*Valeria, Navifusa, Ostiana, Sphaerocongregus, Asperatofilum, Taenitrichoides*, wide ornamented *Siphonophycus, Polytrichoides, Trachytrichoides*, and *Pseudodendron*) are also known from the Kuzha and/or Tyul'ma deposits, and, depending on the accent to some or other their peculiarities, researches dated their origin appearance by the Middle Riphean, or by the post-Yurmata time (Kozlov *et al.*, 1997; Veis *et al.*, 2000). As is stated in the just mentioned our work, the last dating has no alternative from the viewpoint of biostratigraphy, because the Nugush and Kuzha-Tyul'ma morphotypes listed above and abundant at all levels of the Karatau Group are absent in the type sections of the Burzyan and/or Yurmata groups. In terms of geochronological model of the Precambrian microbiota evolution (see in Veis *et al.*, 2001, for details), any other solution of the indicated dilemma would mean that the majority of Karatavian microfossils (those from the Nugush Subformation included) represent transit taxa for the entire Riphean (*Stratigraficheskaya...*, 2000). This would be absolutely inconsistent with the empirical distribution patterns of microfossils in the Riphean stratotype of southern Urals and, as is shown recently (Sergeev and Lee Seong-Joo, 2001; Veis and Vorob'eva, 2002), in its Siberian hypostratotype of the Uchur-Maya region.

Thus, the pre-Nugush regional biostratigraphic level, the earliest one among others considered above, is likely near the boundary between the Yurmata and Karatau groups and puts limit to the downward distribution of Late Riphean microbiota in the Riphean stratotype of southern Urals. It is impossible to define precisely its position in the Yurmata-Karatau transitional interval until the termination of discussion about the character of the Karatau Group lower boundary and its position relative to the Kuzha and Tyul'ma strata bearing microfossils similar to some important Nugush morphotypes. It is reasonable to remember therewith that recent geological, isotopic-geochronological, paleontological, and other data led to mutually exclusive inferences (Larionov, 1994; Kozlov *et al.*, 1991, 1994, 1997; Filippov, 1997, 2000, 2001; Veis *et al.*, 2000; Vinogradov and Gorozhanin, 2002; Maslov *et al.*, 2001).

Four microbiotas from Borehole Kabakovo-62 drilled near the Karatavian stratotype area played a key role in the microphytological subdivision of Upper Riphean deposits in the region. These microbiotas are from the borehole intervals of a very limited thickness corresponding to depth ranges of 5100–5099, 4765–4762, 3639–3636, and 3528–3526 m, i.e., to deposits of the Tukaevo, Ol'khovyi, Priyutovo, and Shikhan formations, respectively (Andreev *et al.*, 1981). According to recent stratigraphic scheme (*Stratigraficheskaya...*, 2000), the Tukaevo and Ol'khovyi formations are subdivisions of the Middle Riphean Serafimovka Group, and the Priyutovo and Shikhan formations are discriminated in the Upper Riphean Abdulino Group. In view of data available at present, their microbiotas can be correlated with the microfossil succession in the Riphean stratotype in the following way. The Tukaevo and Ol'khovyi formations yield single microfossils, remarkable among which are some peculiar *Satka* forms and large lophostriated *Valeria*. In the Bashkirian meganticlinorium, the latter are known from the Bederysh-Uk, Tolpar, and Tyul'ma deposits, whereas the Burzyan and Yurmata groups (the upper Yurmatinian Revet Subformation included) are lacking them. In terms of biostratigraphy, these data are consistent with an earlier inference that the Tukaevo and Ol'khovyi formations bearing *Valeria* belong to the Upper Riphean (Yankauskas, 1982; *Mikrofossilii...*, 1989). They also imply the post-Yurmatinian (post-Revet) age of the earliest biotic events (pre-Nugush and pre-Kabakovo levels) responsible for compositional transformations of microbiotas in the Cis-Ural and southern Urals (Veis *et al.*, 2000).

In addition to a few Riphean transit forms, the younger Priyutovo microbiota includes large *Siphonophycus, Rectia, Brevitrichoides*, and *Polytrichoides* (Andreev *et al.*, 1981). In the Karatavian type section, these morphotypes are characteristic of the Nugush-Uk interval, though without any distinct confinement to particular levels of this broad stratigraphic range. Their distribution does not exclude the formerly suggested correlation between the Priyutovo and Bederysh microbiotas (*Stratotip rifeya...*, 1982) but puts some doubts on its validity, and the problem of the pre-Muldakaevo biotic event in the Cis-Ural region remains open.

The especially representative Shikhan microbiota is terminal one in the succession of microfossils recovered by the borehole under consideration. In addition to all morphotypes of the Priyutovo assemblage, it includes *Tasmanites, Octaedrixium, Spumosina, Trachyhystrichosphaera truncata* Herm et Jank., *T. aimika* Herm., *Palaeolyngbya sphaerocephala* Herm. et Pyl., *Polyspaeroides, Chlorogloeopsis, Tortunema, Glomovertella, Pellicularia*, and *Arctacellularia* (Andreev *et al.*, 1981). In the Upper Riphean type section, most of these forms first appear in the Shishenyak microbiota only. Based on this similarity between the Shishenyak (lower Inzer) and Shikhan microbiotas and on lithologic similarity of the correlative Katav and Shikhan formations (*Stratigraficheskaya...*, 2000), it seems reasonable to assume that the pre-Shishenyak event of microbiota transformation took place somewhat earlier in the Cis-Ural region than in the Karatavian type area. This conclusion exemplifies a phenomenon of appearance of forms-predecessors in the evolution history of Riphean microbiotas. Similar forms have been recognized earlier in some North Siberian sections with

abundant microfossils, in those their parts, which are lacking distinct stratigraphic hiatuses but anticipate the latter (Veis *et al.*, 1998a, 2001).

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The Middle–Upper Paleozoic Stratigraphy and Paleogeography of the Transitional Area between the Turkestan Ocean and Tarim Continent (Tien Shan)

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Abstract—New data on the Devonian–Carboniferous stratigraphy in the western part of the Kokshaal-Tau Ridge elucidate geological history of the transitional area between the former Tarim continental shelf in the south and the Turkestan paleoocean in the north. The conodont zonation (additionally substantiated by data on foraminifers, radiolarians and other fossils) is established in the Famennian–Bashkirian sequences of carbonate-clastic deposits of the continental slope and deep-water calcareous–siliceous sediments. The area under consideration was a large (not less than a few hundreds of kilometers wide) pre-collision area of heterogeneous crust. The continental slope (the Kokshaal unit) was bordered in the north by the Aksai pelagic zone with separate volcanogenic and carbonate buildups. In the Middle Paleozoic, the development of carbonate platforms, slopes and bathyal zones was associated for a long time with intraplate basaltic and contrast volcanism. A transgression of the initial Tournaisian time ceased the avalanche sedimentation of material derived from the Tarim continent and triggered a wide development of bituminous sedimentation. In the late Visean–Serpukhovian, it brought about the carbonate deposition in the pelagic zone. Hiatuses in the Middle Carboniferous bathyal sequences represent most likely the sediment dissolution. Fusulinid successions are used to reconstruct more accurately the southward migration of the flysch trough in response to collision of the Tarim margin with the Kazakhstan paleocontinent during the Bashkirian–Late Carboniferous time.

Key words: Devonian, Carboniferous, pelagic zone, carbonate platforms, conodonts, foraminifers, collision, southern Tien Shan, Tarim, passive margin.

Stratigraphy and lithology of the Middle–Upper Paleozoic deposits in central areas of the southern Tien Shan (the Atbashi–Kokshaal region) can be used to get a deeper insight into the evolution of the Tarim continent passive margin and to consider a complicated paleogeography of the adjacent part of the Turkestan ocean, which separated this paleocontinent from the Paleo-Kazakhstan.

The problem of the Turkestan ocean opening is especially ambiguous in the eastern part, where the reliably dated ophiolites of spreading origin are absent. In the west (the Kyzylkum, Fergana) and east, a mature Atlantic-type stage of the ocean development began with a widespread carbonate sedimentation in the mid-Silurian and terminated in the mid-Carboniferous at the commencement of the collision. The fold-and-nappe structure of the area adjacent to the Turkestan paleobasin was under formation since that moment through the Permian (Biske, 1995, 1996). Amplitudes of thrusting and the current convergence of paleobasin margins indicate that the basin area repeatedly experienced a transverse contraction. Initial positions of the Ulan–Bozoi carbonate platform and analogous structures, and of pelagic areas and volcanic ridges in the Turkestan ocean have been considered earlier (Biske, 2001) based on materials collected by Biske *et al.* (1985). The procedure of paleotectonic reconstructions includes the following operations: recognition of structural units based on their constituent pre-collision formations and characteristic thickness of the latter; determination of time moments, when the units accreted to the Paleo-Kazakhstan margin and turned into tectonic nappes; and an attempt to assess the subsequent nappe movements. A series of original units (nappes) appeared to have been thrust southward, over the Tarim continent.

STRUCTURAL UNITS OF THE WESTERN KOKSHAAL–MAIDANTAG RIDGES AND PRE-COLLISSIONAL EVENTS

The structural units are described below in the north-to-southward succession, i.e., beginning from the upper nappes toward the lower ones, as they cross the western end of the Dzhangdzhir Ridge and the Zapadnyi Aksai River basin, the areas of our investigations in 1998–2000 (Fig. 1).

The Turkestan paleoocean suture. This tectonic element corresponds to the Atbashi–Inyl'chek suture

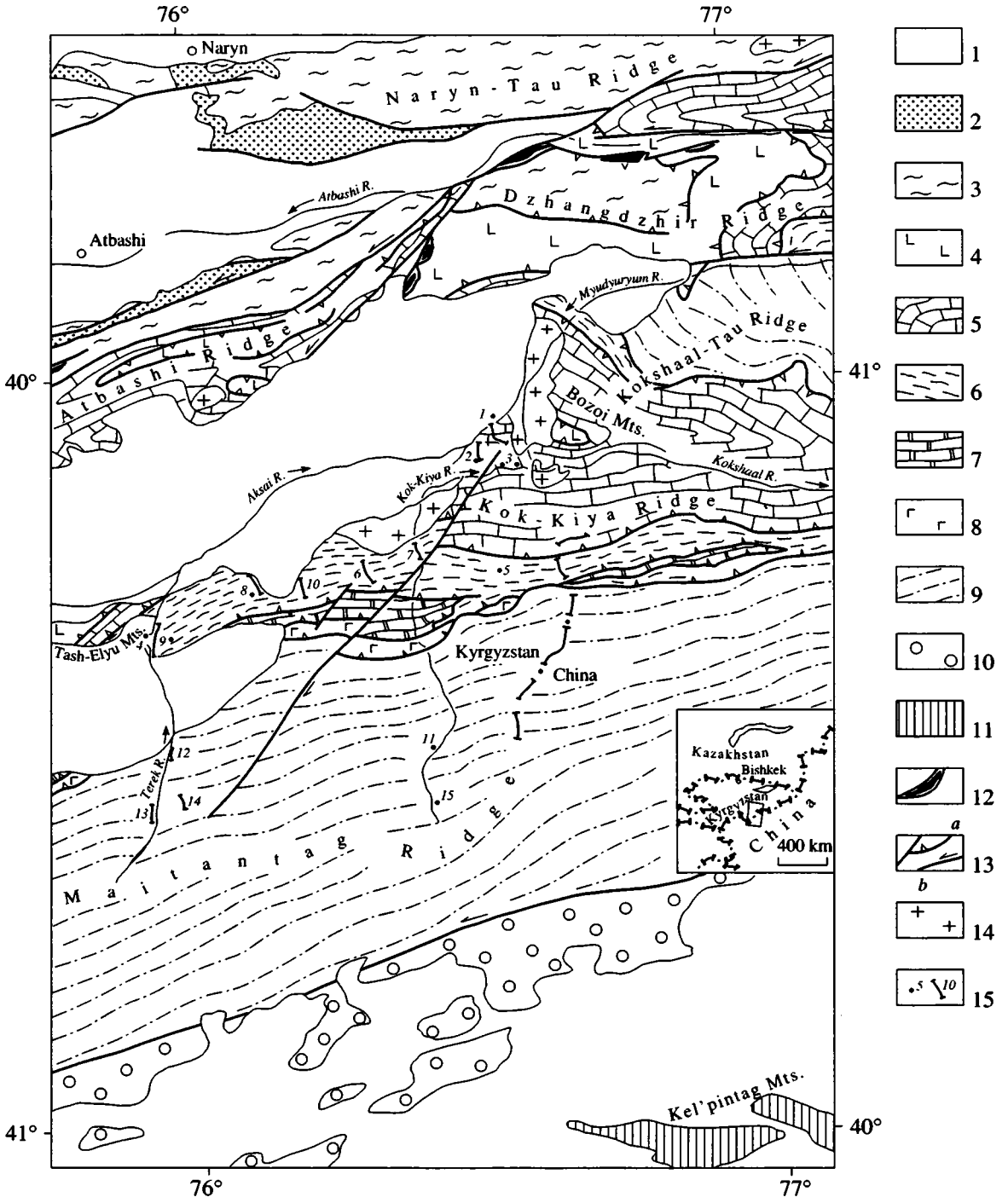

Fig. 1. The Paleozoic structure of a part of the Atbashi–Kokshaal region: (1) Cenozoic cover; (2) rear trough of the Kokshaal Hercynides (volcanogenic–sedimentary Upper Carboniferous and Lower Permian deposits); (3) Paleo-Kazakhstan with the Atbashi and Balykty formations and other metamorphic complexes of the marginal accretionary prism; (4) Dzhanydzher, (5) Ulan, (6) Aksai, (7) Sarybeles, (8) Ortosu, (9) and Kokshaal structural units of the fold-and-thrust structure of southern Tien Shan; (10) Muzduk region (the Upper Paleozoic foredeep and the Middle Paleozoic reefal margin of the Tarim continent); (11) Paleozoic cover of the Tarim platform; (12) ophiolite melange; (13) early collisional thrusts (*a*) and late collisional shears (*b*) with indicated movement directions; (14) collisional granitoids; (15) points and sections referred to in the text and Figs. 2–4.

with adjacent small fragments of serpentinite melange (Khristov and Khristova, 1980; Kurenkov, 1983). The initial suture zone has been significantly modified by the northward retro-thrusts and sinistral shear of the late collision time, and by neotectonic dislocations. The suture zone and allochthonous fragments of tectonized

ophiolites of the left bank of the Atbashi River include slabs and smaller blocks of variously metamorphosed ultramafic rocks, gabbro, gabbro-pegmatite, skarns, and basalts (epidote-chlorite shales, amphibolites). The basalts are intercalated with gray and variegated radiolarites bearing *Entactinosphaera grandis* Naz., *E.* cf. *quantilla* Foremann, and other taxa suggesting the Frasnian age of the upper sedimentary cover above the oceanic crust.[1] The melange is faulted against the overlying turbidites of the Balykty Formation locally transformed into chlorite-sericite schists. The turbidites are referred to the Silurian?–Lower Devonian of the Paleo-Kazakhstan continental slope (Khristov and Mikolaichuk, 1983). Above them there are strongly dislocated Moscovian–Upper Carboniferous deposits age of the neoautochthon.

The Dzhanydzher unit. In the Dzhangdzhir Ridge, the ophiolite melange or the flyschoid Balykty Formation overlies a series of thrust slices composed of Devonian–Lower Carboniferous volcanogenic–siliceous rocks. The slices constitute the Dzhanydzher unit, which may be divided into the upper, predominantly siliceous Borlutor Formation of small thickness and the lower, volcanogenic Kainar Formation of significant thickness. Summarizing previously published data (Biske *et al.*, 1985; Biske and Tabuns, 1991; Biske, 1996, and others), we can outline the following main structural features important for the paleogeographic and geodynamic characterization of the unit.

(1) The unit is composed mostly of the Silurian–Lochkovian graptolitic shales, conodont-bearing radiolarites spanning all other Devonian stages, and Lower Carboniferous micrites and calcarenites, all deposited in the pelagic zone. Some groups of volcanoes and related reefal limestones appeared locally in that zone during the Late Silurian and became more widespread in the Emsian–Eifelian.

(2) Volcanism was of the intraplate character. Subalkaline Ti–Fe-tholeiites are locally intercalated with less abundant rhyolites (the Kainar Formation, Lower–Middle Devonian).

(3) At the top of the unit there is a thin member (a few tens of meters) of coarse flyschoid deposits with local olistoliths. According to its conformable occurrence on Visean–Serpukhovian clastic limestones, the member is not younger than the early Bashkirian.

The first two peculiarities lead to a somewhat controversial interpretation. In terms of geochemistry, the Kainar rhyolites are products of the granite layer contamination in the continental crust. On the other hand, the pelagic environments could not remain stable for such a long time (Silurian through Early Carboniferous) in shelf areas or near large shoals, which supplied clastic and carbonate material. Thus, we should admit that the unit was formed outside the Tarim shelf, where granite layer was likely developed in the crust, like in some recent oceanic rises. The repeated intraplate magmatism of the Late Silurian–Devonian time resulted in a relative depth decrease. After the magma cooling, carbonate platforms were formed on volcanic structures, and avalanches and turbid flows of volcanogenic–calcareous material took place.

The third feature suggests that the Dzhanydzher unit was accreted to the Paleo-Kazakhstan margin in the earliest Middle Carboniferous, when transformation of the Turkestan ocean into the suture zone was probably at the commencement.

The Ulan unit. The unit corresponds to a large nappe complicated by additional internal thrusting in the western Kokshaal Ridge and to the synform of the Bozoi–Kokshaal river valley in the east. The unit is known as the Ulan carbonate platform (Biske, 1996). In the exposed part, it includes the limestone massifs of the Tash-Elyu and Kodzhege–Bozoi Mountains, though in general the limestones are concealed under the Cenozoic sedimentary cover of the Aksai depression.

At the base of the unit, there are Middle Devonian intraplate basalts 600 m thick (Biske and Tabuns, 1996). They are overlain by the Givetian–Bashkirian limestone sequence, which is continuous and up to 4000 m thick in some places and reduced to a few hundred meters in the others (Tash-Elyu) because of elimination of the Devonian–Tournaisian interval. The unit is crowned with a flysch–olistostrome formation of the Bashkirian age.

In the Myudyuryum River lower courses, a lower part of the limestone sequence (Poyarkov, 1969; Khristov, 1970) (up to 2500 m thick) consists of bedded limestones and dolostones, which yield Givetian, Frasnian, and Famennian foraminifers. The facies composition suggests lagoonal environments of carbonate accumulation behind reefs. At least the Famennian interval is mostly composed of oolitic-pelsparite, occasionally fenestrate limestones with a small admixture of crinoid detritus. We studied predominantly the Carboniferous strata in the valleys of the Zapadnyi Aksai River and its tributaries upstream of the Myudyuryum River mouth (Fig. 1, sections 1 and 2). Facies changes in limestone beds exposed in tectonic slices and blocks revealed a transition from shallow-water massive limestones in the south to slope clastic varieties with siliceous admixture in the north (Bozoi Mountains). The latter are subdivided based on conodonts and foraminifers into the following units (Neevin, 2001).

Syphonodella Beds. The lower Tournaisian boundary is approximately at the level of replacement of light-colored Famennian limestones by dark brecciated varieties with members of black bedded cherts or cherty flaggy ("varved") calcarenites, among which there are bioclastic (crinoid) facies. Conodonts are represented by the Tournaisian assemblage of *Syphonodella* cf. *sulcata* (Hudd.), *S.* cf. *isosticha* (Cooper),

[1] Paleontological and age determinations are performed, if not indicated otherwise, by S.M. Likhoman (radiolarians and tentaculites), A.V. Neevin (conodonts), T.Yu. Vorob'ev, A.V. Dzhenchuraeva, and O.F. Getman (foraminifers).

Polygnathus siphonellus Dr., *P. normalis* Mill. et Young., *P. purus purus* Vog., *Bispathodus stabilis* (Br. et M.), *Neopolygnathus communis* (Br. et Mehl.), and *Palmatolepis gracilis sigmoidalis* Zieg. Judging from the rock microstructure, some Devonian conodonts are probably redeposited from Famennian pelsparite beds. Foraminifers are represented by *Parathuramminites cushmani* (Sul.), *Chernyshinella* cf. *disputabilis* Dain, *Bisphaera minima* Lip., and *Tournayella vulgaris* Lip. Thickness of the *Siphonodella* Beds is within the range of 180–250 m.

The typicus Zone (the lower upper Tournaisian). The zone corresponds to thicker beds of oolitic and algal, predominantly dark-colored, frequently breccia-like limestones with cherty nodules. The rocks contain conodonts *Gnathodus typicus* Cooper, *Neopolygnathus communis* (Br. et Mehl.), *Polygnathus purus purus* Vog., radiolarians *Entactinia vulgaris microporata* Braun, *E. tortispina* Orm. et Lane, and foraminifers *Endothyra (Latiendothyra) latispiralis* Lip., *E. (Laxoendothyra)* cf. *paracosvensis* Lip., *Tournayella pigmea* Leb., *Chernyshinella glomiformis* (Lip.) and others. The beds are from 10–20 to 300 m thick.

The anchoralis Zone (the upper upper Tournaisian). Fine- to medium-bedded limestones with interbeds of black platy cherts and cherry–brown siliceous-clayey shales represent rocks of the zone. The upper, predominantly massive oolitic–pelsparite limestones yield abundant foraminifers *Septaglomospiranella micula* Vdov., *Earlandia vulgaris* (Raus. et Reitl.), *Tournayella discoidea* Dain f. *maxima, Glomospira brevispira karakubensis* Vdov., *Granuliferella granulosa* H. Zeller. Conodonts *Scaliognathus* sp., *Protognathodus praedelicatus* Law *et al.*, *Gnathodus semiglaber* (Bisch.), *G. typicus* Cooper, and *Neopolygnathus communis* (Br. et Mehl.) occur throughout the zone interval. Cherts yield radiolarians *Entactinia vulgaris vulgaris* Won. and *E. v. microporata* Braun. Unlike the underlying beds, the zone is characterized by a significant bioclastic admixture: along with foraminifers and algae, there are skeletal detritus of crinoids, bryozoans and others. The zone is from 15–30 to 150 m thick.

The texanus Zone (lower–middle Visean). In the lower part, the limestones are coarsely bedded, oolitic, grading into typical oolitic sparite with pellets, bioclasts (fragments of brachiopod valves and other detritus), and algal remains. The middle part is composed of limestone turbidites having a fine cyclic structure (rhythms of a few centimeters thick), which include interbeds and diagenetic lenses of dark-colored cherts and occasional clay micrites. The zone contains conodonts *Gnathodus texanus* Roundy, *G. semiglaber* (Bisch.), *G. pseudosemiglaber* (Thomp. et Fell.), *Pseudognathodus symmutatus* (Rhod., Aus. et Dr.), foraminifers *Endothyra boumani* Phill., *Septaglomospiranella* cf. *subsymmetricus* Vdov., *Tetrataxis sussaica* Mal., *Omphalotis amplis* (Schlyk.), *O.* cf. *intrequensis* (Schlyk.), *Uralodiscus* cf. *cordailicus* Mal., *Pseudoendothyra* aff. *angulata* (Raus.), etc., and radiolarians *Entactinia vulgaris microporata* Braun, and *Entactinosphaera? trendelli* Won. Thickness range in different sections is from 60 to 390 m.

The bilineatus Zone (the lower part of the upper Visean) and *the overlying, up to the mid-Bashkirian deposits.* Dark flaggy limestone beds range in thickness from a few centimeters to 1 or 2 m. The dominant fine- to medium-grained calcarenites are composed of biogenic detritus and ooliths. Oo- and pelsparite with foraminifers are less frequent. Thicker beds probably represent the storm and avalanche deposits. At their base, there are accumulations of detritus (crinoid columnals, brachiopod valves, rugose fragments). Upward the section the rock particles decrease in size and oolitic varieties are dominant. Occasional tempestite interbeds up to 10 cm thick either are cross-bedded or reveal the overturned microfolds of a landslide origin. All the rocks suggest the carbonate material transport down the paleoslope, i.e., toward the north in modern coordinates. Some hummocky bedding planes (hard grounds) may reflect the long-term breaks in sedimentation. Thinner cyclites show graded bedding. The background sediment in their upper parts is black fine commonly spicule–foraminiferal calcarenite grading into reddish marl. Diagenetic chert nodules occur throughout the section, but true siliciliths are rare.

The largest part (about 800 m) of this interval is exposed near the Kynda River mouth. The rocks yield here conodonts *Gnathodus texanus* Roundy, *G. pseudosemiglaber* Thomp. et Fell., *G. semiglaber* (Bisch.), *Lochriea commutata* (Br. et Mehl.) and foraminifers *Archaediscus convexus* Grozd. et Leb., *Omphalotis* cf. *minima* (Raus et Reitl.), *Priscella prisca* (Raus. et Reitl.), *Eostaffella mosquensis* Viss., *E.* aff. *rotunda* Durk., *E. settella* Gan., *Howchinia gibba* (Moell.), *Parastaffella struvei supressa* (Schlyk.), *Paraarchaediscus koktjubensis* (Raus.), *Pojarkovella honesta* Sim., *Globoendothyra* sp., *Endostaffella shamordini* (Raus.) and others. Species *Lochriea commutata* implies that this part corresponds to the late Visean (beginning from the *bilineatus* Zone) through, probably, the Serpukhovian stages. The latter date is inferable from occurrence of the Serpukhovian *Eostaffella* ex gr. *protvae* Raus. in association with *Neoarchaediscus gregorii* Dain and other species of this genus and with goniatites *Cravenoceras?* sp. and *Dimorphoceras?* sp. (identifications of L.A. Ektova and A.V. Yagovkin), all found in the same deposits of the Bozoi area (communication of V.L. Klishevich, 1975). The uppermost limestone beds of the Kynda River section do not differ in lithology from the underlying beds and contain early Bashkirian *Pseudostaffella* ex gr. *antiqua* Dutk. The same beds of the Bozoi area yield Bashkirian conodonts (communication of V.P. Chernyshuk, 1991). The thickness of the Kynda section is 850 m, and the complete succession, the Bashkirian Stage included, is estimated to be as thick as 1000 m.

The Bashkirian limestones are conformably overlain by a member of marls and shales (up to 25–50 m) representing a deep-water "pre-flysch". Above them there is the Konurtyube Formation of terrigenous–carbonate turbidites with limestone olistoplaques and of gravel–pebble deposits of a gravitative type, which contain carbonate, siliceous and basaltic clasts. Presumably reworked foraminifers from this formation exposed at the Kok-Kiya River (Fig. 1, point 3) are of the Serpukhovian age. Chernyshuk and Gushchin (unpublished data, 1991) discovered that the formation included an olistoplaque with foraminifers of the *Plectostaffella bogdanovkensis* Zone (the base of the Bashkirian).

The data presented above allow us to reconstruct the history of the Ulan carbonate platform in more detail. Its mature stage has been reached by the end of the Devonian. A transgression that started in the early Tournaisian produced zones of uncompensated subsidence (siliciliths), which later turned into channels or straits partly filled with reworked carbonates. A sedimentation rate was relatively low in the Tournaisian–early Visean, though not less than 40–50 m/m.y. in the late Visean, as is evident from the deposits thickness. It is possible to suggest that the upper Visean–lower Bashkirian sediments described above accumulated in the upper part of a wide and gentle slope of the carbonate platform before reefs rather than in a shelf area. In the Ulan shallow-water zones (e.g., the Tash-Elyu Mountains in the Aksai upper reaches), there are upper Visean–Serpukhovian reef limestones overlying transgressively the Devonian deposits. Taking into consideration the direction of sediment transport, we admit a southward progradation of the platform slope in the study area.

The tectonic subsidence of the basin floor and the incomplete compensation continued in the Tournaisian through early Bashkirian. The process ceased in response to the collision event that is recorded by deposition of the flyschoid Konurtyube Formation in connection with the northward advancing and destruction of upper nappes in the inner (volcanogenic) zone of the Kokshaal Hercynides.

The Aksai unit. The tectonic slices exposed at the right bank of the Zapadnyi Aksai River are composed of Middle Paleozoic bathyal deposits, which can be divided into three distinct groups.

The lower group is represented by dark graptolite shales of the upper Llandovery–Lochkovian. The Wenlockian age of deposits has not been proved by fossils. The *spineus-formosus* Zone of the upper Ludlow and the Pridoli to Lochkovian deposits are fully represented, being up to 200 m thick in total. The background clay sediments are intercalated with distal carbonate turbidites and calcarenite interbeds. The rocks bear volcanomictic and quartz (terrigenous?) admixture. The deposits are analogous of "the Kursala Formation" of the Alai Ridge (Koren' and Lytochkin, 1992). More proximal facies occasionally of avalanche

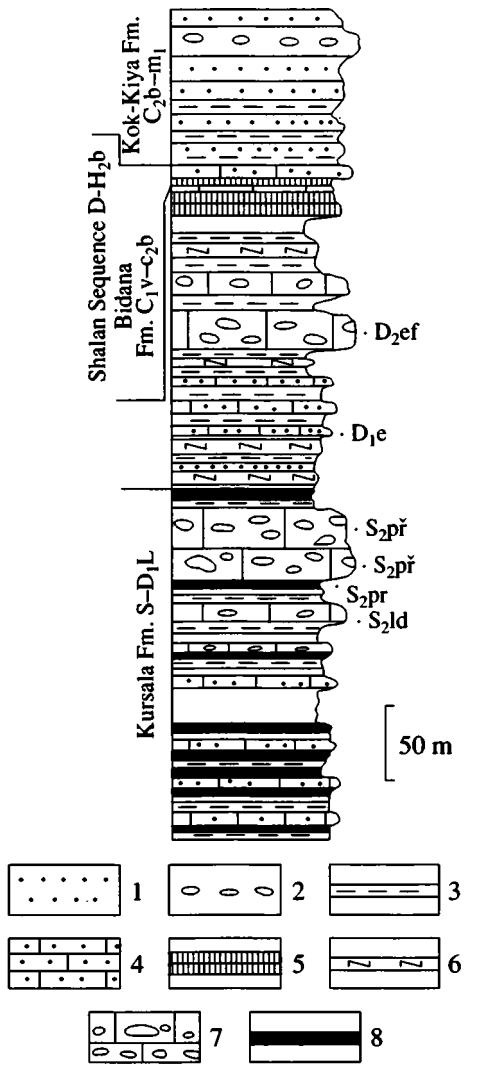

Fig. 2. Lithostratigraphy of the Aksai structural unit with frequent olistostromes (Fig. 1, point 4): (1) graywacke; (2) conglomerate; (3) aleuropelite; (4) calcarenite; (5) black silicilith; (6) red and green silicilith (jasper); (7) calcirudites, blocky ones included; (8) dark aleuropelite (graptolitic shale); geologic indices based on fossil evidence (see the text) are designated to the right of the column.

origin, which appear in some sections (Fig. 2; Fig. 1, point 4), are represented by carbonate breccias and other calcirudites with reef limestone clasts containing the Ludlovian corals *Propora* sp. and the Pridolian assemblage with *Squameofavosites bohemicus singularis* Sok., *Sq.* cf. *ettkychuensis* Chekh. and other forms. The background carbonaceous shales yield graptolites, the Pridolian *Pristiograptus* ex gr. *transgrediens* Perner included. Thus, the clastic material was transported from evolving limestone buildups.

The next, most complete group of predominantly siliceous deposits is of a wide stratigraphic range, from the Pragian through basal Bashkirian. It is analogous in general to the Shalan Group of the Turkestan-Alai

region (*Stratifitsirovannye...*, 1982) but shows in distinction a variable amount of volcanic and carbonate intercalations. Its group thickness ranging from 170 to 400 m suggests the low sedimentation rate of an "oceanic" type (2–5 m/m.y.). The lower unit of the group (the Surash Formation) consists of siliciliths of dingy green to pinkish coloration owing to the variable admixture of carbonate, silty–clayey, and fine-grained sandy materials. These rocks are referred to the Devonian based on rare (in spite of closely spaced sampling) Emsian conodonts *Polygnathus* aff. *gronbergi* Kl. et John. Like in the Silurian sequence, some sections (most frequently along the Terek River lower course, Fig. 3) reveal presence of avalanche breccias of reefal limestones (up to 50 m thick). The latter contain crinoid detritus (*Cupressocrinites*) and Emsian tabulates (different forms of *Favosites, Oculipora, Gracilopora*, and *Yacutiopora* identified by Biske). The upper member yields *Caliapora* aff. *uralica* Yanet, brachiopods *Zdimir pseudobaschkiricus* Tchern. (identification by Savitskii), and other Eifelian benthic fossils. More distal facies of the Pragian–Emsian clastic carbonates were discovered further to the south, upstream the Terek River. They are represented by finely bedded micrites and siltstones bearing dacryoconarids *Turkestanella acuaria* (Richt.), *Paranowakia* cf. *scalaria* Schloth., *Nowakia* cf. *cancellata* Richt., and conodonts *Polygnathus filiformis* Snig., *Pandorinellina steinhornensis miae* (Bult.) and others. When admixture of clastic carbonate material is significant, the Lower Devonian–Eifelian deposits can be as thick as 350 m.

The carbonate platforms of the pre-collisional basin evolved partly on the volcanogenic substratum. This is evident from basaltic flows (up to 150–200 m) intercalating with siliceous beds at the right bank of the Kok-Kiya River. The flows are composed of subalkaline Ti-basalts of the intraplate type (Biske and Tabuns, 1996). The latter are overlain by radiolarites with Frasnian *Entactinia prodigialis* Naz., *E.* cf. *consociata* Naz., *E. tenuiacerosa* Naz., and *Entactinosphaera variacanthina* For. (Fig. 1, point 5). In the majority of sections studied, the Givetian–Frasnian age of deposits has not been proven by fossils, although this interval may correspond to silty siliciliths and micritic limestones of a small thickness (up to 60 m).

Predominantly dark color of the upper Frasnian, Famennian and Tournaisian deposits is a consequence of repeated (?) anoxic conditions during this time interval. The deposits constitute two formations of different lithology.

The lower, Kul'dzhabashi Formation is readily recognized in the Kul'dzhabashi and Kok-Kiya river valleys (Fig. 1, sections 6, 7), where it is composed of black to greenish silty–clayey shales and siliciliths with interbeds of calcarenites (in places reworked oolitic sparite) bearing admixture of basic volcanic clasts. Some interbeds correspond to storm deposits, and the other ones with graded bedding are undoubtedly proximal turbidites of the carbonate platform. The limestone detritus contains an assemblage of foraminifers and microalgae. Foraminifers represented by abundant archaesphaers *Parathurammina tuberculata* Lip., *P. paulis* Byk., *Avroria singularis* Pojark., *A. ferganensis* Pojark., *Quasiendothyra* sp., *Bisphaera malevkensis* Bir., *Radiosphaera* sp., *Kamaena* sp., *Calcifolium plavkensis* Reitl., and *Quasiumbella pararhotunda* Shv. are associated with radiolarians *Entactinosphaera grandis* Naz., *E.* cf. *egindyensis* Naz., *E. foveolata* Naz., and *E. diversita* Naz., and with conodonts *Polygnathus symmetricus* E.R. Branson. The cherts yield Famennian conodonts *Palmatolepis inflexoidea* Zieg., *P. glabra* Ull. et Bas., *P.* cf. *minuta* Br. et Mehl. and the same species of radiolarians. The formation is 100–110 m or commonly less thick depending on variable amount of clastic material in sediments. For instance, the Famennian interval (identified based on conodonts) in the Terek River section is represented only by dark red and green silty siliciliths.

The upper, Urusai Formation is distinguished by predominantly siliceous composition and dark coloration. Limestone interbeds concentrated in its upper part contain reworked foraminiferal detritus and abundant Tournaisian radiolarians *Entactinia vulgaris* Won., *Entactinosphaera foremanae* Orm. et Lane, *E. egindyensis* Naz., *E. crasseacus* Lich., *E.* cf. *aitpaiensis* Naz., *Eostylodictya eccentrica* Orm. et Lane, *Trilonche cimelia* Orm. et Lane, *Cromiostylus variabilis* Orm. et Lane, and *Provisocyntra* sp. associated with conodonts *Polygnathus spicatus* E.R. Branson, *P. purus* Vog., *Bispathodus stabilis* (Br. et Mehl.), *Protognathodus?* sp., and *Dollymae hassi* Vog. Two last forms from cherts of the Ichke-Kyzylsu River section (Fig. 1, section 8) are indicative of the lower upper Tournaisian. The black Urusai cherts are usually not more than 30–70 m thick.

The Visean–Serpukhovian and lower Bashkirian deposits are represented by flaggy cherty calcarenites and micrites with spicules and radiolarians, which are occasionally intercalated with clay interbeds or substituted by bedded siliciliths. These deposits widespread in southern Tien Shan have been described as the Bidana Formation (*Stratifitsirovannye...*, 1982). The lower part of the formation yields calcareous algae *Masloviporidium* sp. (identification by S.B. Gushchin) known from in the Venev Horizon of the Russian platform. This suggests a stable carbonate sedimentation in a bathyal zone of the late Visean basin. The foraminiferal assemblage from the lower beds (20–25 m) of the formation proves the above inference and consists of abundant *Eostaffella* ex gr. *postmosquensis* Kir., *E.* sp., *Cribrostomum bradyi* Moell., *Tuberitina maljavkini* Mikh., *Semistaffella?* sp. and reworked *Archaeosphaera* and *Quasiendothyra*. The limestones also yield two conodont assemblages. The lower Visean–Serpukhovian assemblage includes *Gnathodus bilineatus bilineatus* (Roundy), *G. b. bollandensis* (Hig. et Bouck.), *Lochriea nodosa* (Bisch.), *L. commutata* (Br. et Mehl.), *L.* cf. *cruciformis* (Clarke), *L.* cf. *mononо-*

dosa (Aus., Rhod. et Dr.), *Pseudognathodus symmutatus* (Rh., Aus. et Dr.), and *Ps. homopunctatus* (Zieg.). The assemblage from upper 25–30 meters consists of *Declinognathodus noduliferus* (Ell. et Grov.), *D. inaequalis* Hig., *Neolochriea hisaharui* Miz., *Idiognathoides sulcatus* Hig. et Bouck., and *Id. fossatus* (Br. et M.) characteristic of the lower Bashkirian. The Bidana Formation is not more than 100–120 m in total, but the Lower Carboniferous interval is almost missing in some sections (e.g., in the Terek River lower reaches, Fig.1, section 9).

The upper sedimentary sequence of the Aksai tectonic slices is represented by the Kok-Kiya Formation (Biske *et al.*, 1985), which is composed of a proximal sandy flysch. The main rock types are calcareous graywackes with a minor admixture of volcanogenic clasts (basalts). They frequently occur as thick (up to 1–3 m) grain flows separated by siltstone interbeds. Gravel and pebbles in some gravitite beds (up to 7 m thick) and in small olistoliths are composed of underlying limestones (those with Devonian fossils included), siltstones, Upper Devonian siliciliths, less frequent porphyrite, and intrusive aplite. Presence of plant detritus and fragments of large horsetail stems indicate that sedimentary material was derived from a land that emerged as "thrust cordilleras" composed of bottom sediments of the former deep basin. However, the intercalated calcirudite lenses originated probably by deposition of material derived from shoals of a narrow unstable shelf area and from older dislocated limestones (Lower Carboniferous, according to foraminiferal records). The unique limestone breccias and sandstones (400 m thick) at the Ichke-Kyzylsu River (Fig. 1, section 6), which contains clasts with different-aged (Famennian to early Moscovian) foraminifers, must be of similar origin and age.

As one can infer from oldest foraminifers found and reworked conodonts, the Kok-Kiya Formation was deposited in the Bashkirian–Moscovian interval, from the *Ps. antiqua* or, more likely, *Ps. praegorski* time to the Kashirian time. In some sections, the formation is up to 400–450 m thick. The thickness is surprisingly small for the corresponding type of sedimentation during the long interval. In the Tekelik River section (Fig. 1, section 10), the formation begins with calcarenite interbeds containing Bashkirian foraminifers of the *Pseudostaffella antiqua* Zone (250 m). In other sections, the basal beds of the formation contain early Moscovian *Profusulinella* and *Aljutovella* forms. Under conditions of the accretion prism growth, there were probably the short-term episodes of sedimentation. As a result, the Bashkirian deposits rest conformably on the Lower Carboniferous limestones, whereas the Moscovian strata overlie the erosional surface of the latter.

According to the above characteristics, the sedimentary complex of the Aksai unit is typical in general of Middle Paleozoic pelagic sequences of insignificant thickness, which accumulated in the bathyal zone of the southern Tien Shan. The described Middle Paleozoic complex of the Dzhanydzher unit (except for the volcanogenic deposits) is of the same type. Below is our interpretation of some (particular and general) structural features of the Paleozoic Aksai sedimentary complex.

(1) A significant though not common admixture of reworked calcareous material in the Pridoli–Eifelian deposits was derived from reefs and carbonate platforms, which evolved above the intraplate volcanic buildups. These could be either the limestone massifs of the Ulan and Sarybeles units or other buildups concealed under nappes or denudated later.

(2) The sporadic occurrence of Givetian and Frasnian deposits and the reduced thickness of Upper Devonian bathyal sediments, while the carbonate accumulation in shelf areas and large shoals was intense, may be indicative of the basin floor subsidence below the carbonate compensation depth.

(3) The dark coloration (bitumenization) of bathyal and shelf Tournaisian facies everywhere in the study region (carbonate rocks of the Central Tien Shan, Alai, Borkoldoi, and other microcontinents of southern Tien Shan included) may be the result of the stagnant regime development in oceanic bottom waters.

(4) The replacement of pelagic siliceous facies by limestones, which took place almost everywhere by the end of the Visean, might be caused by a regional, if not by a global event. Besides the evident transgression, which is recorded in shelf sequences of the South Tien Shan and nearby continents, the increased avalanching of slope sediments, which was characteristic of the Ulan carbonate platform, may point to the lowering the carbonate compensation depth in the basin, although the bottom waters were better aerated than in the Tournaisian.

(5) The delayed deposition of the flysch cover (the Kok-Kiya Formation) in the Aksai unit, as compared to similar events in the Dzhanydzher and Ulan structural units, can be easily explained in terms of successive, from the north to the south, movement and accretion (collision?) of thrust nappes of the latter units. A considerable time span of the process (Bashkirian to mid-Moscovian) is indicative of a large width of the Aksai pelagic area, if we assume a steady-state convergence of the colliding plates.

The Sarybeles and Ortosu units. These two tectonic nappes located at present to the south of the Aksai unit were described earlier (Biske, 1979, 1996). The Kel'tau–Sarybeles limestone succession recorded the formation of a stable carbonate platform that had no connections with that of the Ulan unit. Taking into account all the exposures of analogous limestones, including those in China, we may conclude that the platform was not less than 100 km long. A volcanogenic buildup is fragmentally exposed below the Wenlock–Ludlovian cover of organogenic carbonates. In

the Kok-Kiya River valley, shallow-water massive limestones of the Lower–Middle Devonian laterally grade into slope calcirudites and calcareous–clayey-siliceous bathyal sediments.

In the Givetian time, basaltic eruptions resulted in emergence of the Ortosu seamount and signified the commencement of a new subsidence cycle that was associated with organogenic limestone deposition on the adjacent Sarybeles shoal. In the Early Carboniferous, especially in the late Visean, the Sarybeles unit was crowned by transgressive limestone horizons analogous of a thin reefal member of the Ortosu unit. After deceleration and cessation of tectonic subsidence, the early Moscovian transgression resulted in deposition of the fusulinid limestone member. The lower Moscovian turbidite–olistostrome deposits at the top of the Sarybeles section were deposited, when the carbonate shoal was sinking into the deep trench in front of the Paleo-Kazakhstan active margin. The Sarybeles–Ortosu shoal was separated from the Tarim continental slope by the deposition setting of Devonian bathyal radiolarites of the Aksai type, which are now exposed immediately southward.

The Kokshaal unit, or the Yassy–Maidantag zone. The thick Uzgen sequence of Devonian sandy turbidites deposited in the Tarim continental slope is a distinctive feature of the unit. At least in the eastern Kokshaal zone, these rocks rest on Silurian–Lower Devonian shelf sediments, i.e., they overlie the continental crust (Biske *et al.*, 1992). The structure of the Kokshaal unit is well observable in the right-side area of the Zapadnyi Aksai riverhead. The unit is composed here of several tectonic sheets thrust southward. Large folds corrupted by different-amplitude shears complicate the reconstruction of complete stratigraphic succession and the assessment of rock thickness. The difficulties were partly overcome by means of detailed transects in several valleys.

Only an upper part (the Tysbel Formation) of the Devonian terrigenous deposits is exposed in northern slopes of the Maidantag Mountains. The formation is mainly composed of sorted greenish oligomictic sandstones. They constitute both the rhythmic flysch members and the sandstone cyclites a few meters thick, which reveal large rod- and tongue-like mechanoglyphs at their basal planes. The cyclites alternate with thin silty-clayey interbeds characterizing the background sedimentation. The cross bedding orientation in sandstones suggests the northward movement of the flows. The formation is supposed to be formed by the material discharge from grain flows, which transported the mature, predominantly quartz clastics from the weakly eroded continent. The deposition occurred in an avandelta of a gentle accumulative continental slope. The minimal thickness of the formation is about 2000 m. Paleontological materials support the above interpretation. The silt interbeds contain accumulations of terrestrial plant detritus derived from different places (Late Devonian *Pseudobornia ursina* Nath., *Sphenopteris* sp., *Svalbardia* sp., *Pteridorachis* sp., and *Rhacophyton* sp., identifications of 1979 by N.M. Petrosyan). They also yield abundant conodont fragments, including the late Famennian *Palmatolepis gracilis* Br. et M., and *Siphonodella* sp., which were found by Biske at the Korumda River (Fig. 1, point 11) and identified by Chernyshuk. These species suggest the Tournaisian age of the uppermost part of the formation. Carbonates are usually insignificant components of the turbidites, though there are sandy calcarenites with foraminiferal tests of the Famennian–Tournaisian age. Red siltstones occur at the top.

The northern margin of the Kokshaal zone hosts more pelagic facies of the Famennian deposits, at least in the upper part of the sequence. These are mainly greenish and variegated thin beds of silty–clayey turbidites, which are intercalated with rare sandy and calcarenite-sandy members. The Famennian age is indicated by foraminifers · *Eotuberitina* aff. *ferganensis* Pojark., *Tournayella* sp. and others.

These deposits are overlain by the carbonate–siliceous Karasainyn Formation of the Lower–Middle Devonian. The lower part of the formation consists of radiolarites, which are black, less frequently gray–green in color, and of slightly bituminous silty–clayey or occasional sandy rocks. The formation is 40 to 50 m thick in general, being as thick as 200 m in a single place. The siliciliths yield the conodont assemblage of *Siphonodella sulcata* (Hud.), *S. duplicata* (Br. et M.), *S. cooperi* Hass, *S. isosticha* (Coop.), *Bispathodus stabilis* (Br. et M.), and *Polygnathus* ex gr. *inornatus* Br. et M., which are indicative of the early Tournaisian. The age determination is frequently consistent with occurrence of radiolarians *Entactinosphaera aitpaiensis* Naz., *E. foremanae* Orm. et Lane, *E. egindyensis* Naz., *Thecentactinia riedeli* (Foreman), *Cromyostylus variabilis* Orm. et Lane, *Paronaella turgida* Orm. et Lane, *Eostylodictya eccentrica* Orm. et Lane, *Trilonche cimelia* Orm. et Lane, *Archocytrium clinoceras* Deflandre, and *Albailella indensis ambigua* Braun.

The black siliceous member of the lower Tournaisian is exposed everywhere in the Kokshaal unit (Biske *et al.*, 1992). It marks an interruption of sedimentary material avalanching from the Tarim continent probably because of a transgression that brought about the stagnant warm-water hydrodynamic regime.

The upper Tournaisian deposits (20–40 m, occasionally up to 120 m), which have been discovered in 11 sections studied, are of a similar lithologic composition. In the Terek section near the Ortosu River mouth (Fig. 1, section 12, Fig. 4), the upper Tournaisian siliciliths span the *typicus* and *anchoralis* zones (Biske and Neevin, 2000). The thicker *typicus* Zone contains conodonts *Dollymae hassi* Vog., *Protognathodus praedelicatus* Lane, Sand. et Zieg., *Pr.* cf. *cordiformis* Lane, Sand. et Zieg., *Pseudopolygnathus* ex gr. *triangulus* Vog., *Neopolygnathus communis* (Br. et Mehl.), *Polyg-*

nathus purus Vog., *Gnathodus typicus* Coop., and *G.* aff. *delicatus* (Br. et Mehl.). The *anchoralis* Zone yields *Scaliognathus anchoralis* Br. et M., *Gnathodus delicatus* Br. et M., *G. pseudosemiglaber* Thom. et Fell., *Hindeodella segaformis* Bisch. and some forms of wider stratigraphic ranges, all associated with radiolarians *Entactinia vulgaris microporata* Braun, *E. v. vulgaris* Won, *E.* cf. *brilonepsis* Won, *E. tortispina* Orm. et Lane, and *Entactinosphaera aitpaiensis crasseacus* Lih. msc.

It is remarkable that the extremely low rate of deep-water sedimentation that is characteristic of the Maidantag Ridge sections is much higher in the eastern Kokshaal zone, where thick turbidites related to the Tarim provenance prevail over the black siliceous sediments (Biske *et al.*, 1992; Biske, 1996).

The Visean green to cherry red clayey shales and siliciliths differ coloration from the Tournaisian sediments and occasionally yield abundant conodonts. The latter are most frequently represented by an assemblage of the *texanus* Zone. The assemblage includes *Gnathodus texanus* Roundy, *G. girtyi girtyi* Hass, *G.* ex gr. *girtyi* Hass, *G. semiglaber* (Bisch.), *G.* cf. *pseudosemiglaber* Thomp. et Fell., *Pseudognathodus symmutatus* (Aus., Rh. et Dr.), and *Ps. homopunctatus* (Zieg.). Conodonts of the *bilineatus* Zone established less confidently and frequently are *Lochriea commutata* (Br. et M.), *Gnathodus* ex gr. *girtyi* Hass, *G. bilineatus* (Roundy), *Pseudognathodus homopunctatus* (Zieg.), and *Lochriea* cf. *multinodosa* (Wirth). Thickness of both zones is about 20 m (Fig. 1, section 12 at the Terek River), being greater (80–100 m) only if the reworked carbonate material is present in sections as calcarenite and micrite beds.

The Serpukhovian–Bashkirian interval (beginning from the *nodosa* Zone) predominantly corresponds to calcareous beds (fine-bedded, frequently clayey micrites and thin partly cherty calcarenites) of the same red–green coloration. Conodonts are well distinguishable on their bedding planes. The beds 40 to 50 m thick in total locally bear sandy admixture. The largest part of the interval corresponds to the *bollandensis* Zone of upper Serpukhovian. The Serpukhovian goniatites *Karanchoceras* sp., *Cravenoceras?* sp., and *Tympanoceras* sp. (identifications of 1976 by A.V. Yagovkin), the half-dissolved shells of which occur frequently in clay limestones, probably belong to this zone. *Gnathodus postbilineatus* Nigm. et Nemir. found in a single place suggests presence of the uppermost Serpukhovian *postbilineatus* Zone (Fig. 3). The Bashkirian deposits (not more than 30–40 m thick) yield scarce fossils of the *corrugatus* Zone, characteristic of which are *Idiognathoides corrugatus* Har. et Holl., *Id.* cf. *sinuatus* Har. et Hall., *Id. sulcatus* Hig. et Bouck., *Id. fossatus* (Br. et Mehl.) and rare *Reticuloceras* sp. We failed to recognize the lowermost Bashkirian *noduliferus* Zone of the conodont standard. Its absence may be indicative of a break in sedimentation and/or of the sediment dissolution. In the most complete Kokshaal sections, the overall thickness of the Karasainyn Formation is 80–250 m.

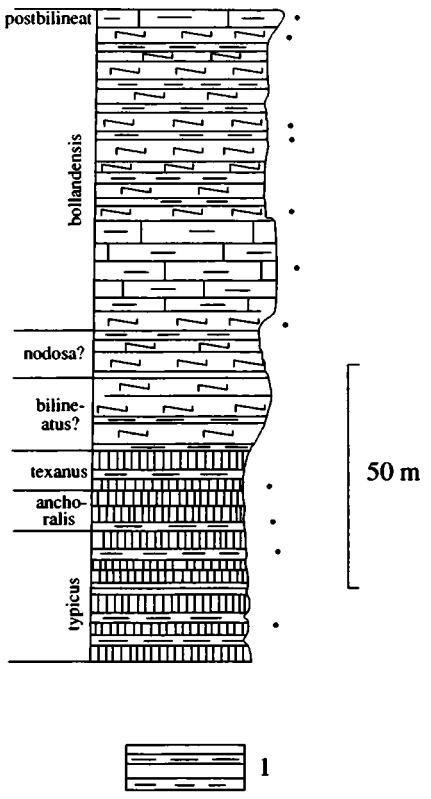

Fig. 3. Lithostratigraphy of bathyal deposits of the Kokshaal unit (Karasainyn Formation, Lower–Middle? Carboniferous); conodont zonation and sites of representative sampling (points to the right of the column) are shown: (1) clay limestone; other symbols as in Fig. 2.

At the top of the Kokshaal unit there is the Kipchak Group of the early collision flysch, which presumably originated, when a of residual bathyal part of the continental rise subducted under the imbricated accretionary complex, and, correspondingly, a foredeep begun to develop instead of the oceanic trench.

The Kipchak flysch is structurally similar to the older Kok-Kiya Formation described above. The terrigenous sandy material is represented mostly by quartz, silicilith, and limestone clasts brought from the northern areas. The flysch conformably lies on the bathyal (Karasainyn) deposits or locally, in the northern part of the Kokshaal unit, it is above a transitional pre-flysch member. The member is composed of alternating siliciliths and distal turbidites, including the calcareous ones. The calcarenites contain the foraminiferal assemblage with *Pseudostaffella* ex gr. *antiqua* Dutk. and *Profusulinella* sp., (identifications of L.A. Ektova). The associated conodonts *Idiognathoides sinuatus* Har. et Holl., *Id. corrugatus* (Har. et Holl.), and *Id. sulcatus* Hig. et Bouck suggest the late Bashkirian–early Moscovian age of the member. The overlying flysch contains fossils not younger than early Moscovian.

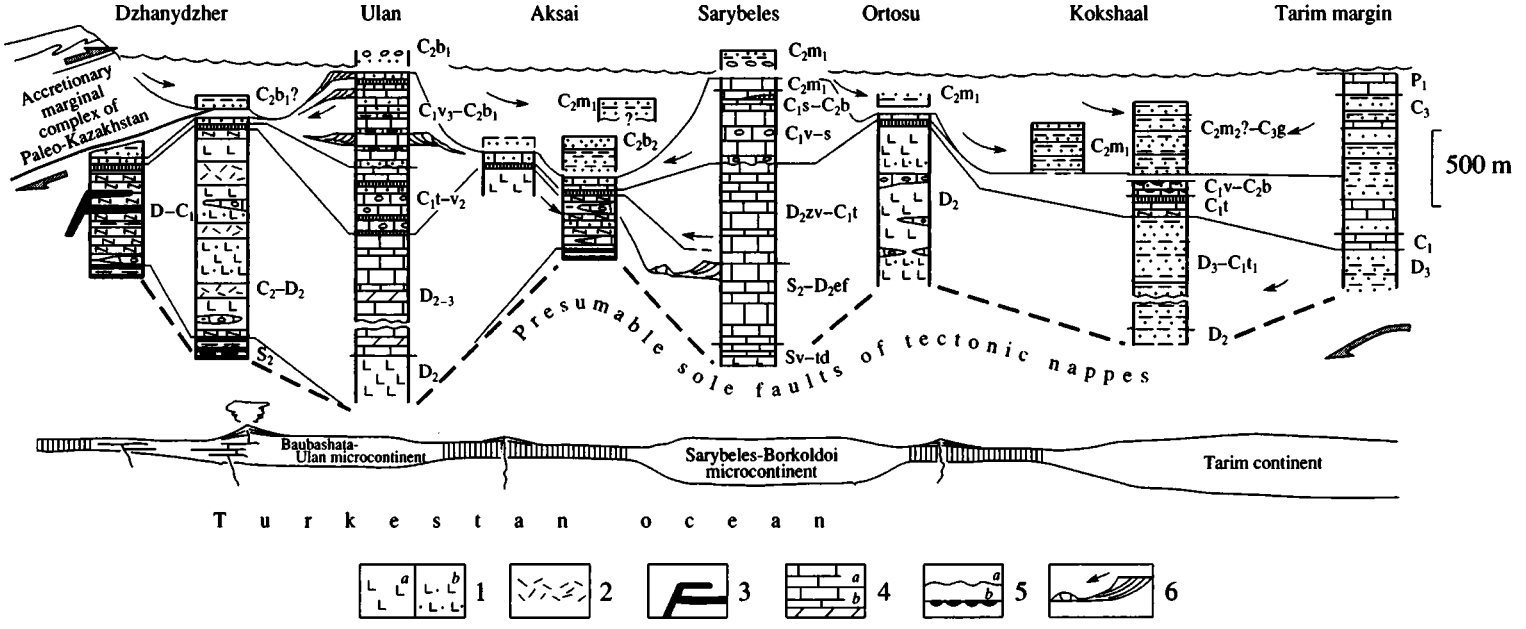

Fig. 4. Structural model of the Tarim northern margin in the western Kokshaal sector at the commencement of collision: (1) basic lavas (*a*) or hyaloclastites (*b*); (2) acid volcanic rocks; (3) intrusive dolerite sills; (4) shelf limestones (*a*) and dolostones (*b*); (5) breaks in shallow-water (*a*) and deep-water (*b*) sedimentation; (6) direction of sedimentary transport; clinoforms and olistostromes are shown schematically, other symbols as in Figs. 2 and 3. Shown below the wavy line (sea level) is approximate configuration of the basin floor at the beginning of the syncollisional formation of flysch (column parts above the bottom). Thick arrows indicate directions of the rock mass movement during the collision. Presumable zones of oceanic (vertical hatching) and continental crust are shown in the sketch below columns of Paleozoic deposits.

In the southern near-boundary part of the Kokshaal Ridge, the Kipchak Group is of a different structure. The basal beds of calcareous sandstones contain here *Fusulina mosquensis* Raus., *F. quasifusulinoides* Raus., *F.* cf. *kamensis* Saf., *Fusulinella* ex gr. *colaniae* Lee et Chen, *F.* cf. *fluxa* Lee et Chen, *F. pseudobocki ovoides* Raus, *F.* aff. *altispiralis* Bog., *F.* aff. *helenae* Raus. (Fig. 1, points 13, 14). Above them, the rocks yield *Obsoletes* cf. *vinutus* Kir., *O.* aff. *pauper* Vol., *O. dagmarae* Kir., *Protriticites* cf. *lamellosus* Bensch, *P.* cf. *paramontiparus* Raus., *P. variabilis* Bensch, and *P.* cf. *subvatus* Bensch. *Triticites* ex gr. *expressus* Anos. is found further upward. According to this succession of fossils, the sequence may be referred to the upper Moscovian–Gzelian, but the evidently reworked state of foraminiferal tests suggests the Gzelian age of its largest part. In any case, the southern Kokshaal sections lack a considerable part of the Moscovian Stage. To explain this, we suggest a balance between the pelagic sedimentation and submarine leaching of calcareous material before the flysch deposition.

The Kipchak Group encloses olistoplaques of Lower Carboniferous limestones and siliciliths. In a single place (the upper reaches of the Korumda River: Fig. 1, point 15), the flysch includes a 10-m-thick member of black radiolarites with conodonts *Gondolella elegantula* Stauf. and *Idiognathodus* sp. (identifications of Chernyshuk) characteristic of the upper Moscovian or Upper Carboniferous. This indicates that the deep-water pelagic environments existed in the Kokshaal trough at least until the end of the Moscovian.

To complete the paleogeographic picture (Fig. 4), it should be noted that outlines of the Tarim continental shelf and southern (outer) limits of the foredeep are reconstructed based on the Muzduk area sections, the southern Maidantag Ridge slope and the Kel'pintag Ridge (Biske, 1996; Zhang *et al.*, 1983; *The Carboniferous...*, 1987). In the Middle–Late Carboniferous, sedimentary material was transported into the southern Maidantag–Kokshaal foredeep from the south, i.e., from the Tarim shelf (Carroll *et al.*, 1995). The conclusion is consistent with deposition of clastic limestones (carbonate turbidites) of this age in the Kipchak bathyal zone, but siliciclastic material could be brought there from the north only. It is possible that some Kipchak sections of the boundary (watershed) area are composed of materials derived from both provenances. In the mid-Early Permian, the foredeep experienced the collisional deformations, which were followed by the late, continental stage of the foredeep development.

MAIN RESULTS

In the Middle–Late Silurian, the northern margin of the Tarim paleocontinent and the transitional area between it and the Turkestan paleoocean, including the area studied, represented a mosaic of relative uplifts (carbonate platforms) and extensive pelagic areas.

In the Middle Paleozoic, the lithosphere cooling, thickening and subsidence typical of the passive margins were complicated by the intraplate magmatism of the Late Silurian–Middle Devonian. The continental crust fragments occurred even in the pelagic zones, as it is evident from the contrast composition of volcanic series. At that time, there were local accumulations of carbonate and volcanogenic clasts (olistoliths) in the pelagic zones, where the uncompensated sedimentation persisted, however, up to the mid-Carboniferous. This indirect evidence suggests a considerable width of the former open basin that became many times shorter during development of the observable fold-and-nappe structure of the region.

The detailed Tournaisian–lower Bashkirian stratigraphy of the continental slope, pelagic (bathyal) zones, and carbonate platforms is based on conodont succession additionally controlled by distribution of radiolarian and foraminiferal assemblages. The inferable major events of that time are as follows: cessation of sediment avalanching from the western Tarim margin; widespread development of anoxic conditions (bituminous sediments) and sedimentation rate decrease on the carbonate platforms in response the early Tournaisian transgression; progradation of the carbonate platforms and their slopes with the concurrent preservation of calcareous sedimentary aprons in pelagic zones of the late Visean–Serpukhovian time; and partial dissolution of older carbonate oozes that eliminated some lower Bashkirian and Moscovian strata.

The early Bashkirian collisional events started in the Dzhangdzhir area advanced southward during the Middle–Late Carboniferous and successively affected the inner marginal zones of the Tarim paleocontinent, producing the thrust–nappe structures. The stratigraphic successions are dated predominantly based on fusulinids. All the sections demonstrate the asynchronous conformable transition from pelagic sediments to flysch that is indicative of the Tarim plate subduction followed by appearance of an oceanic trench, which was rapidly filled with material eroded from the accretionary prism. In the Permian, the extreme southern part of the trench was transformed into the Kokshaal foredeep. The Late Paleozoic history of the Kokshaal zone reconstructed above is similar to successions of collisional events in many linear orogens.

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In Defense of the Volgian Stage

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Abstract—The Volgian Stage was a subdivision used in Russian and world geological literature for over a century. The stage was withdrawn from the general scale and replaced by the Tithonian Stage, when the Interdepartmental Stratigraphic Committee of the Russian Federation (ISC RF) came to that decision on February 2, 1996. The paper casts doubts on the decision. A return to the use of parallel stages in the International stratigraphic scale for the time intervals with high geographic differentiation of biota is substantiated.

Key words: Volgian stage, biostratigraphic zonations, parallel stages, Panboreal Paleobiogeographic Superrealm.

The Volgian Stage, the uppermost one in the Jurassic Stratigraphy of boreal sediments, was in use over a century in theoretical and applied works of Russian and foreign geologists. As it was recommended by different-rank International stratigraphic organizations, the Volgian Stage figured during 27 years (1964 to 1991) in the International stratigraphic scale autonomously or in parallel with the Tithonian Stage (Ager, 1964, 1974; *Rekomendatsii mezhdunarodnogo simpoziuma...*, 1974; Saks, 1979). At the meeting of 1991 held in Potie, France, the International Subcommission on the Jurassic System decided to withdraw the Volgian Stage from the general stage scale (Cope, 1993). By ISC RF decision of February 2, 1996, the Volgian Stage was divided into the Jurassic and Cretaceous parts, withdrawn from the general scale, and replaced by the Tithonian Stage (Zhamoida and Prozorovskaya, 1997; Rostovtsev and Prozorovskii, 1997). Is the century-long history of the Volgian Stage, the uppermost one of the Boreal Jurassic, terminated? Are the Volgian Stage potentialities exhausted? Is the problem of its correlation with the Tithonian Stage solved? Is the Tithonian Stage identifiable, as easily as the Volgian Stage, over the entire distribution territory of boreal sediments? These questions are not rhetorical. Despite the above-mentioned decision, the Volgian (not Tithonian!) Stage is still used in publications devoted to stratigraphy of Northern Europe, northern North America, all Arctic islands, the huge Greenland inclusive, and shelves of the Barents, Kara, Laptev, East Siberian, Chukchi, and Beaufort seas (figure). Such a practice is lawful, because the stage, in the term meaning suggested by d'Orbigny, corresponds to a group of beds with specific fossils (Cope, 1996). In other words, every stage can be recognized based on its characteristic fauna. It is logical therefore that the Volgian Stage was established based on foraminifers even in Madagascar (Kuznetsova, 1976). According to the figurative sentence by Egoyan (1984), the stage is recognizable owing to its "stiffening core." Ammonites are the orthostratigraphic fossil group for the Jurassic System and even for the entire Mesozoic. It is principal that ammonite "cores" of the Tithonian and Volgian stages are substantially different: it is impossible to find even a single coincidence at the species and generic levels in the nomenclature of over 30 constituting zones and subzones of these stages.

Adherents of the Tithonian Stage applicability to the Boreal Realm are dazzled by idea of the unified nomenclature of stratigraphic units in the International geochronological scale. It is clear that the stage nomenclature in the general Phanerozoic scale should be unified, but without the loss of its merits. Russian geologists defined the Volgian Stage in stratigraphic successions over the vast territory of northern Eurasia. The Volgian Stage is rendered in hundreds of large-scale maps. It is described and mentioned in thousands of publications and hand-written initial reports since the 19th century until present. The refusal from the Volgian Stage complicated substantially the work for Russian geologists, particularly for those engaged in petroleum geology, whose studies are almost entirely dedicated to the boreal-type sections. Is it reasonable to destroy traditions without any forcible arguments? Globalization is useful when it does not violate national interests.

VOLGIAN STAGE AND THE JURASSIC–CRETACEOUS BOUNDARY

The ISC RF accepted the most radical version of the Jurassic–Cretaceous boundary among all others discussed in relevant publications. It placed the boundary at the base of the upper Volgian Substage (the base of the *Kachpurites fulgens* Zone), and four zones of the upper Volgian Substage turn out to be presumably corresponding to the Berriasian *Berriasella jacobi* and,

Table 1. Ammonite zonations in the Jurassic–Cretaceous boundary interval of European Russia, northern Siberia, and southern Europe

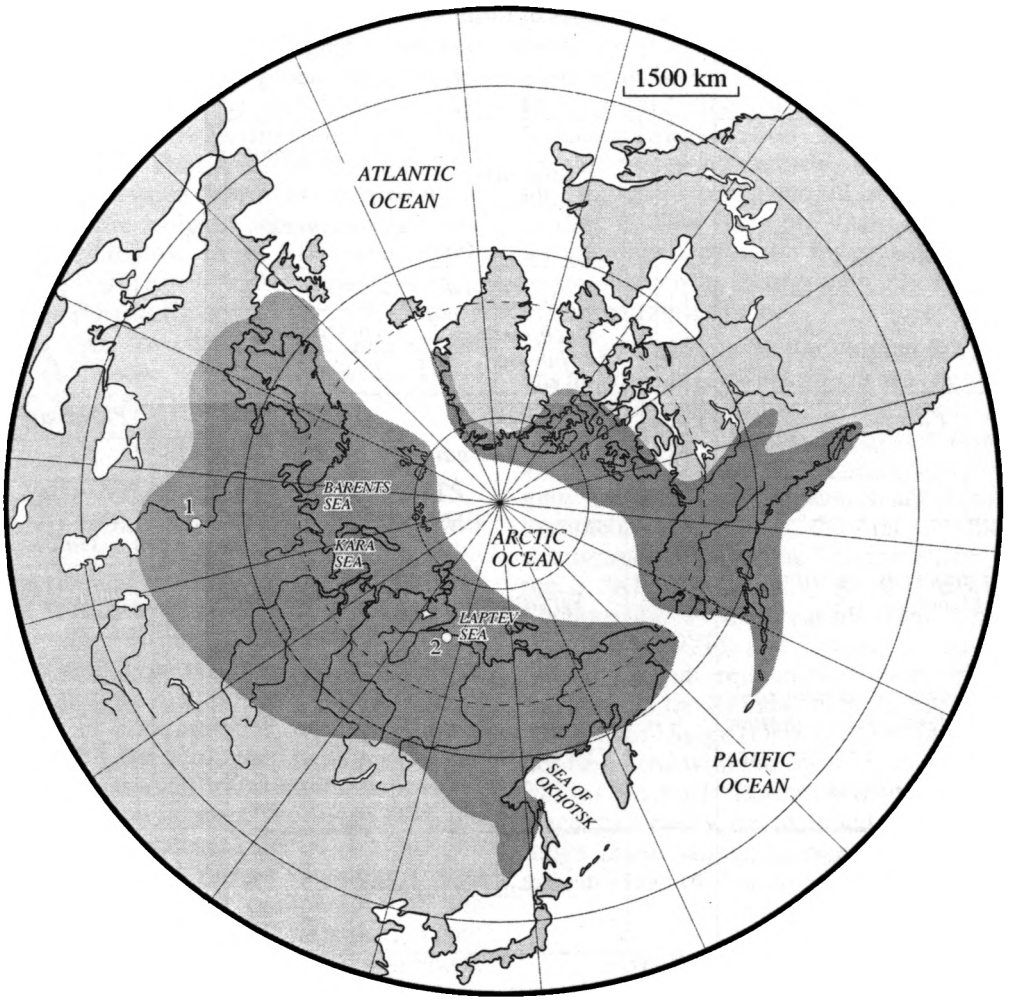
Stage	Substage	European Russia (Mesezhnikov *et al.*, 1979)		Northern Siberia (Zakharov *et al.*, 1997)		Substage	Stage	Stage	Substage	Southern Europe (Hoedemaeker and Rawson, 2000)	
"Ryazanian"	upper	*Peregrinoceras albidum*		*Tollia tolli*		upper	Boreal Berriasian	Berriasian	upper	*Fauriella boissieri*	*T. alpinensis*
				Bojarkia mesezhnikowi							*B.picteti*
		Surites tzikwinianus		*Surites analogus*	*S. analogus*						*Malbosiceras paramimounum*
					S. subquadralus						
	lower	*Riazanites rjasanensis*	*Euthymiceras transfigurabilis*	*Hectoroceras kochi*	*S. praeanalogus*	lower					
			Hectoroceras kochi		*Bor. constans*						
					H. kochi						
			Garniericeras	*Chetaites sibiricus*	*Ch. sibiricus*				middle	*Tirnovella occitanica*	*D. dalmasi*
					Pr. maynci						
Volgian	upper	*Craspedites nodiger*	*C. nodiger*	*Chetaites chetae*		upper	Volgian				*B. privasensis*
			C. mosquensis	*Craspedites taimyrensis*							*T. subalpina*
		Craspedites subditus		*Craspedites okensis*	*C. originalis*				lower	*Berriasella jacobi*	
					C. okensis						
		Kashpurites fulgens	*C. nekrassovi*		*P. exoticus*						
			K. fulgens								
	m.	*Epivirgatites nikitini*		*Epivirgatites variabilis*		m.		T.	u.	*Durangites* spp.	

Note: Solid horizontal line shows one of two versions of placing the lover boundary of the Cretaceous System proposed for official discussion (at the base of the *Tirnovella occitanica* Zone). In case that this boundary is adopted, the Volgian Stage remains almost completely in the Jurassic System. The subdivision of the zone into subzones is after Baraboshkin (1999), (m.) middle; (u.) upper; (T.) Tithonian.

partially, *Tirnovella occitanica* zones (*T. subalpina* and *B. privasensis* subzones). In this case, it is possible to speak about similarity in succession of zones in sections, but not about coincidence of their stratigraphic ranges. Thus, the suggested correlation of Boreal and Tethyan biostratigraphic units is very arbitrary (Sei and Kalacheva, 1993; Zhamoida and Prozorovskaya, 1997). Generally speaking, the decision to divide the Volgian Stage in two parts and to include the upper one into the Berriasian Stage was incompetent, because, like at present, the Global Stratotype Section and Point for the Berriasian lower boundary, i.e., for the base of the Cretaceous System, was not determined in 1996, and, thus, the official upper boundary of the Jurassic System is still unknown. At the meeting of 1995 held in Brussels, the International Working Group on the Berriasian Stage and the Jurassic–Cretaceous boundary recommended to place the base of the Berriasian Stage at the base of three subdivisions: the *Berriasella jacobi* Zone, *Tirnovella subalpina* Subzone, or *T. occitanica* Zone established in sections of southeastern France or Spain (Zakharov *et al.*, 1996). Thus, none of variants of correlation between the Tithonian, Berriasian, and Volgian stages is unable to give an idea of their ranges and of stratigraphic position of the upper Volgian Substage until the upper boundary of the Jurassic system is determined. Moreover, if the lower boundary of the Berriasian Stage would be at the base of the *T. occitanica* Zone, the *Berriasella jacobi* Zone should be included into the Tithonian, i.e., into the Jurassic System, and the Volgian Stage of the latter appears to be without the *Craspedites nodiger* Zone, the only one, which Mesezhnikov (1984, 1989) proposed to transfer into the Cretaceous System (Table 1). Thus, the decision of the ICS RF was evidently premature.

THE VOLGIAN STAGE IN THE GENRAL STRATIGRAPHIC SCALE

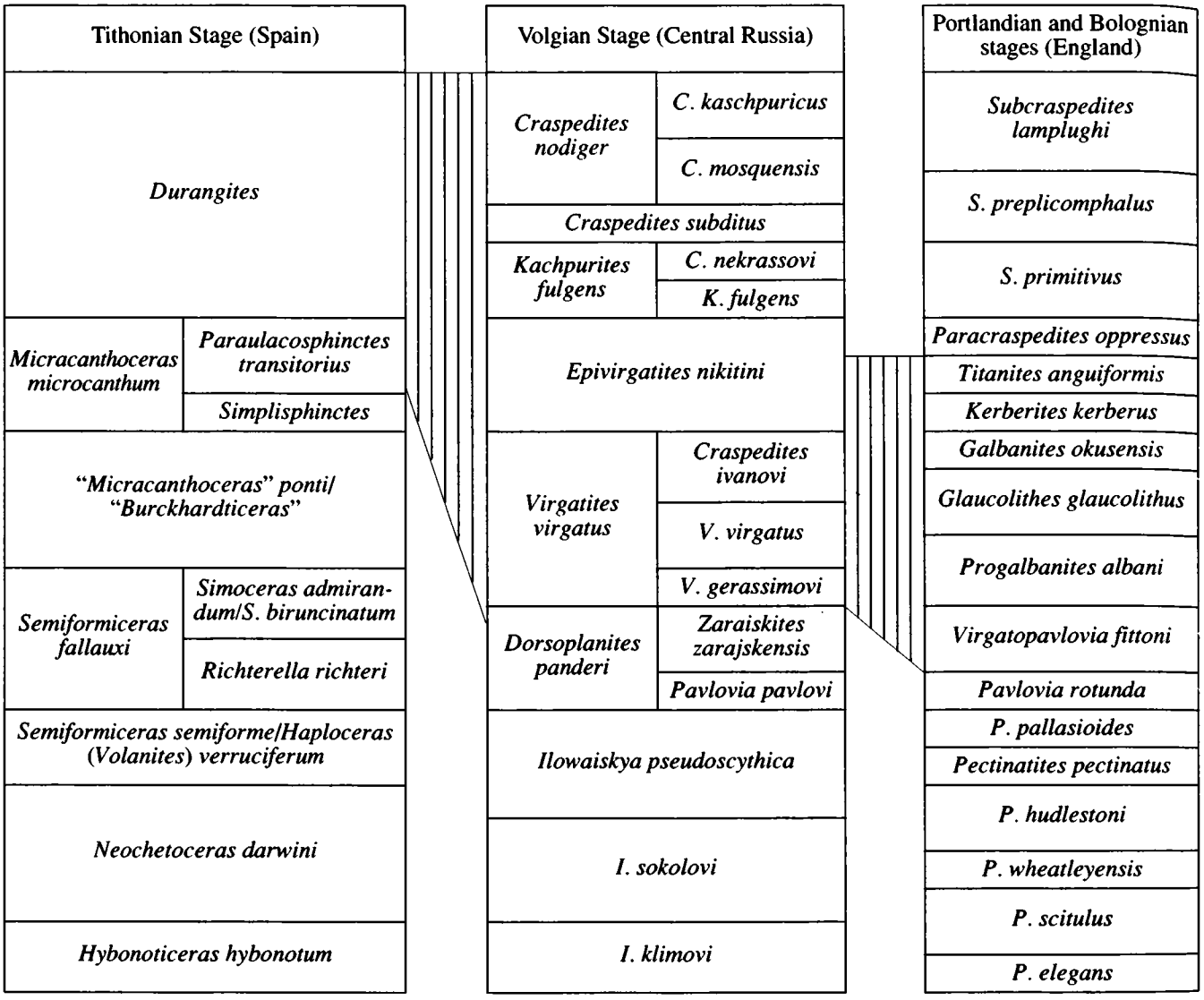
The adjective "Volgian" appeared first in the geological formation name at the end of 19th century, and later it was used to name the Lower Volgian and Upper Volgian stages (Nikitin, 1881, 1884). Both stages corresponded in stratigraphic ranges to the middle and upper substages of the Volgian Stage in its present-day meaning. The lower Volgian Substage as an equivalent of the Vetlayansk Horizon (Sokolov, 1901) was included into the Volgian Stage substantially later (Gerasimov and Mikhailov, 1966; *Postanovlenie rasshirennogo zasedaniya...*, 1966). The decision of



Factual and inferable distribution areas (shaded) of the Volgian Stage sediments in the Northern Hemisphere of the Earth (over 20 million of square kilometers in total). Numbered are sections, the candidates for the GSSPs of the Volgian substage boundaries: (1) Gorodishche site (Ul'yanovsk region), the lower boundary of the Volgian Stage and lower boundary of the middle Volgian Substage; (2) Nordvik Peninsula (Yakutia, Republic Sakha), the lower boundary of the upper Volgian Substage and lower boundary of the Boreal Berriasian.

the ISC SSSR (October 29, 1964) was preceded by decision of the Britain Committee on Mesozoic (February of 1963) to refuse from the Portland Stage in favor of the Volgian Stage and by a later decision of the Committee on the Mediterranean Mesozoic in Cassis (France, May of 1994), which admitted the Volgian Stage validity to replace the Tithonian Stage in the standard scale (Ager, 1964, 1974; Krymholtz, 1974). Nevertheless, the International Symposium on Upper Jurassic Stratigraphy held in Moscow recommended to preserve both the Tithonian and Volgian stages in the general scale (*Rekomedatsii mezhdunarodnogo simpoziuma...*, 1974). Ten years after the Moscow symposium, the International Colloquium of 1977 (USSR: Novosibirsk, Ul'yanovsk, Tyumen, Leningrad), authorized again both stages in the general scale. The aspiration to preserve the Volgian Stage in the general scale was not dictated by the whish of Russian (Soviet) specialists only. To a significant extent, the initiative belonged to *western* paleontologists and stratigraphers. The intense geological studies in northern territories of North America, which were initiated after the World War II in Canada, the United States, Greenland, Svalbard, and, later on, in the Arctic shelf areas, showed that boreal sediments widespread in these region occupy an area over 20 millions of square kilometers in the Northern Hemisphere (figure). Throughout this territory, the Mesozoic sections enclose remains of boreal organisms, those of the Volgian Stage included, which are characteristic of the East European platform. The Volgian Stage started to be distinguished in territories north of the 50°N. The fact noted by K.F. Roulliet (1845) as early as in the middle of the 19th century and later confirmed by works of many Russian geologists (G.A. Trautschold, I.I. Lahusen, S.N. Nikitin, A.P. Pavlov, K.O. Milashevich, A.O. Mikhal'skii, D.N. Sokolov,

Table 2. Ammonite zonations in Jurassic terminal stages of Europe (Tithonian, Volgian, Portlandian, and Bolognian)



Note: Vertical lines designate intervals lacking zones (scales after Cope, 1963; Geyssant, 1997; Baraboshkin, 1993; Mitta, 1993; correlation of the Tithonian and Volgian stages after Rogov, 2002).

A.N. Rozanov, D.I. Ilovaiskii, and others), became evident for geologists of the 20th century. The fact is that paleontological characteristics and composition of the uppermost Jurassic deposits, which represent the Volgian Stage in northern areas of the Northern Hemisphere, greatly differ from their southern counterparts, i.e., from deposits of the Portlandian Stage defined by A. d'Orbigny in the Anglo-Parisian basin (Cope, 1996) and of the Tithonian Stage in Mediterranean areas (Oppel, 1865). The difference is so great that the stages cannot be correlated between each other and their nomenclature should be individual. Nowadays, the "faunal cores" of the stages in question are sufficiently well known. They sharply differ first in taxonomic composition of index ammonite species. As is seen in Table 2, the ammonite assemblages in zonal units of the Volgian and Tithonian stages lack any genera in common. Calpionellids (tintinids), which are important for the Tithonian biostratigraphy, are unknown from the Volgian Stage. To the contrary, buchiid bivalves important for the Volgian Stage biostratigraphy are virtually absent in the Tithonian sediments. Tithonian and Volgian foraminiferal assemblages also have little in common (Ivanova, 1973; Kuznetsova, 1979). Most of marine invertebrates and known vertebrates, which dwelt in the Central Russian basin, were endemic taxa (Gerasimov, 1955, 1969, 1992; Gerasimov *et al.*, 1995). The number of coexisting boreal and Tethyan mollusks increases only in the ecotone zones between 45 and 55°N (Zakharov and Rogov, 2003).

The integrity of the Volgian Stage throughout its distribution area is primarily based on the succession in

development (phylogeny) of characteristic faunal groups. The phylogenetic lineages of higher ammonite taxa from families Perisphinctidae (genera *Ilowaiskiya, Dorsoplanites*, and *Epivirgatites*) and Craspeditidae (genera *Kachpurites* and *Craspedites*) reconstructed within the biogeographic Panboreal Superrealm confirm genetic relation between representatives of these groups during the entire Volgian Age (Ilovaiskii and Florenskii, 1941; Mikhailov, 1966; Shul'gina, 1969; Mesezhnikov *et al.*, 1984; Mitta, 1993). Several species of the boreal genus *Buchia* (bivalves), as well as many other invertebrate taxa are characteristic only of the Volgian Stage (Lahusen, 1888; Zakharov, 1981; Gerasimov *et al.*, 1995; and others).

TITHONIAN AND VOLGIAN STAGES: MERITS, DEMERITS, AND NOMENCLATURE PROBLEMS

Analyzing merits and demerits of the Volgian and Tithonian stages, it is difficult to decide which one has more advantages. The Tithonian Stage is not ideal, because it has no stratotype (Oppel, 1865). Although the problem of stratotype selection is not an exigency after reorientation of International working groups toward GSSP selection, the geological substance is still a main carrier of geohistorical information that cannot be ignored according to sound sense. In this respect, the Volgian Stage with its stratotype sections along the Volga (near Ul'yanovsk) and Yatriya (Subpolar Urals) rivers is more advantageous as compared with the Tithonian Stage (Gerasimov and Mikhailov, 1966; Zakharov and Mesezhnikov, 1974; Mesezhnikov, 1984).

Moreover, the ammonite zonation of the Tithonian Stage is less detailed than that of the Volgian Stage (9 versus 15 zones and subzones). Several autonomous zonations have been elaborated for the Volgian Stage. For instance, the buchiid zonation enables a complete Panboreal correlation and Boreal-Tethyan correlation of lower and middle parts of the stage (Zakharov, 1981). No equivalent zonation is available for the Tithonian. Calpionellids occurring in Tethyan sections only are useful for subdivision of the upper Tithonian sediments only (Remane, 1990). Application of standards based on other fossil groups (benthic mollusks, belemnites, brachiopods, foraminifers, dinocysts) is also limited by distribution areas of different-type sediments.

Although the Tithonian deposits are of significantly wider geographic range (from the Western Mediterranean to Southeast Asia and form the Caribbean basin to South America) than those of the Volgian Stage, they are untraceable north of 50°N in the Northern Hemisphere, and it is also difficult to define them in many areas of the Southern Hemisphere. Even the correlation potential of Tithonian ammonite zones is usually limited by the western Mediterranean region.

The upper Tithonian is missing from the historical locality (southern Germany) and its uppermost *Durangites* Zone is absent in the Berriasian type area. In this respect, the Volgian Stage is equally vulnerable, because sedimentation break and, probably, stratigraphic hiatus are recorded near its top. More exactly, the stratigraphic gap should be most likely at the base of the "Ryazan Horizon" (Mitta, 2001). The break in sedimentation near the lower Berriasian boundary is recorded even in stratigraphically continuous sections of southeastern Spain.

As compared with the Volgian Stage, the Tithonian has an advantage of priority: it was defined by A. Oppel almost 20 years earlier. Nevertheless, as is shown by Mitta (2001, p. 25), the nomenclature stability can be violated, if the variant of correlation of the upper Volgian Substage with a part of the Berriasian will be accepted: "in this case, the first suitable name for the stratigraphic unit (of the Volgian Stage, *V.Z.*) should be the "Khoroshevian" Stage distinguished by Shchurovkii (1867) earlier than the "Berriasian" (Coquand, 1971), and further ... the middle Volgian Substage is an equivalent of the "Moscovian Stage" established by Romanovskii in 1856, i.e., before the "Tithonian" Stage (Oppel, 1865), but the former should not be mixed with the Moscovian Stage of the Carboniferous System (Nikitin, 1890). The lower Volgian Substage can easily be included into the Kimmeridgian Stage sensu anglico or can be defined as the Vetlyankian Stage of Sokolov (1901)."

THE SCALE OF PARALLEL STAGES

The convenience and effectiveness in regional geological works are main arguments in favor of parallel stage scales. Three regional stages (Portlandian, Tithonian, and Volgian) consist of unique successions of ammonite-based zones (Mesezhnikov, 1982; Krymholtz, 1988; Mitta, 1993; Baraboshkin, 1999). Despite undoubted achievements in correlation of the lower parts of these successions (Rogov, 2002), a reliable "through" level suitable for Boreal-Tethyan correlation in the interval from the upper Tithonian to the top of the Berriasian does not exist. The Volgian Stage is well traceable in North Europe (Russia and Poland), Northeast Asia, North America (northern and northwestern Canada, northern Alaska), in all Arctic islands and Arctic sea shelves, that of the North Sea inclusive. The Portlandian Stage is recognizable in the Anglo-Parisian Basin only. Contrary to the opinion of Cope (1996), it is more reasonable to include Greenland in the "sphere of influence" of the Volgian Stage because, in addition to analogues of most zones of the Volgian Stage, established here is the complete succession of buchiid zones, which enable the reliable Panboreal correlation of all three substages of the Volgian Stage (Zakharov, 1981; Hakansson *et al.*, 1981; Surlyk and Zakharov, 1982; Callomon and Birkelund, 1983). The ammonite successions of the Tithonian Stage substantiate correlation of

corresponding Tethyan deposits throughout their distribution area. The parallel calpionellid and nannofossil scales are useful for high-precision correlation of the Jurassic–Cretaceous boundary strata on both sides of the Atlantic (Bralower, 1990; Remane, 1998).

The benefit of parallel stages (and constituting zonal scales) is not limited by determining aspects of stratigraphic correlation. The reliable tracing of genetically similar events is important for the detailed reconstruction of geological history in separate segments of the earth crust. For instance, the Mesozoic geodynamic evolution of the Arctic was controlled by tectonic history of major plates in the Northern Hemisphere, but main biotic events were concentrated in the paleogeographic Panboreal Superrealm, where the Boreal zonations are most efficient (Zakharov *et al.*, 1997). Precisely the Boreal (not Tethyan) regional chronostratigraphy, that of the Volgian Stage included, offers a possibility to trace all the peculiarities stages in formation and evolution of the Arctic Ocean and near-polar regions of the Earth (Zakharov *et al.*, 2002a, 2002b).

Another example is connected with the oil and gas prospecting. It is known that productivity of the West Siberian hydrocarbon-bearing province depends on the generation potential of the high-carbonaceous Bazhenov (Volgian) Formation (Braduchan *et al.*, 1986). Concurrent analogues of this formation are established in shelves of the Kara and Barents seas based on finds of Volgian fossils. These sediments are also bituminous. The molluscan remains of the Volgian Stage are found also in the North Sea deposits with a high C_{org} contents. Commercial oil-and-gas deposits are discovered in all these regions.

The high-resolution stratigraphic schemes are necessary to reconstruct regional paleoclimatic events, to define trends in sedimentation (within paleobasins), and to decipher history of paleolandscapes and paleobiome evolution. Finally, there is no need to use the stratigraphic scale of the Tithonian Stage for comparison of any Jurassic events in the East European, West Siberian, and Siberian platforms, because all these structures were located in the biogeographic Panboreal Superrealm and, consequently, hosted similar biotas.

Thus, the parallel stages of the general stratigraphic scale reflect the real specifics of geological and biological evolution of large segments of the earth crust. They demonstrate peculiar and varied events that occurred in the Earth history. In this respect, terminal regional units of the Jurassic System, such as the Portlandian, Volgian, and Tithonian stages will always be demanded by the routine geological practice (Callomon, 1979). Traditions, which reflect not only terminological convenience and semantic comfort, but also an unambiguous understanding of the problem by many generations of geologists, are an important argument in favor of regional stages.

Recent revision of the International stratigraphic scale by the world stratigraphic community and practice of selecting the GSSPs for stage and substage boundaries demand resolution conflicts of a national character. The experience of International subcommissions on stratigraphic systems shows that their members are far from altruism when selecting a GSSP. The acuteness of debates on the standard is frequently determined by motives of the national prestige, rather than by scientific objectivity. Some authoritative specialists and myself believe that the return to the practice of parallel stages, i.e., of secondary stages in terms of Cope (1993, 1995, 1996) or Abbink and Callomon (2001) can be a temporary measure on the way to unification (Zakharov, 2003) and may help to avoid conflicting situations when selecting GSSPs, making the procedure more objective. Simultaneously, this will also deprive of advantages the voting members of international subcommissions on stratigraphy, which represent the hosting and invited countries (Cope, 1996).

I propose to use the parallel stages along with the main general scale. The introduction of parallel standards is a temporary measure valid until their reliable correlation with the general scale. A parallel standard proposed by national stratigraphic surveys (for instance by ISC RF) must be approved by decision of the International Commission on Stratigraphy (ICS). Boundaries in the parallel stage standard should be determined in the same manner as in the general scale, i.e., by selection of the Secondary Stratotype Section and Point (SSSP). It seems reasonable to place the SSSP for the Volgian lower boundary at the base of the *Ilowaiskiya klimovi* ammonite zone (the base of the lower Volgian Substage) in the Volgian Stage stratotype section near the Gorodishche Village (Volga River near Ul'yanovsk). The SSSP for the lower boundary of the middle Volgian Substage is well defined at the base of the *Dorsoplanites panderi* ammonite zone in the same section. The SSSP of the lower boundary of the upper Volgian Substage should correspond to the base of the *Craspedites okensis* Zone (*Praechetaites exoticus* Subzone) in the Volgian Stage section of the Nordvik Peninsula, Anabar Bay, the Laptev Sea coast (Zakharov *et al.*, 1983).

CONCLUSION

The paper is not aimed to revise views on the problem of the Volgian Stage. I just want to attract attention to one of the most acute problems of the Mesozoic stratigraphy in Russia–to destiny of the Volgian Stage. I doubt the expediency of the ISC RF decision to divide the Volgian Stage in two units, to include the entire upper Volgian Substage into the Berriasian Stage and Cretaceous System, and to withdraw the Volgian Stage from the general stratigraphic scale of Russia imparting to it a status of the regional stage (=horizon). Bearing in mind the fact that boreal sediments cover 90% of the Russian territory, one can easily understand that this formal decision leads to a chaos in work of practical geologists and to unnecessary troubles for specialists

from scientific organizations of the Russian Academy of Sciences. Ignoring the opinion of International subcommissions on stratigraphic systems, the ISC RF came to its decision ahead of recommendations of the International Commission on Stratigraphy (Zakharov *et al.*, 1996). Under motto of unification (who argues?), the ISC RF creates comfortable conditions for some (specialists on Tethyan deposits) and difficulties for others (geologists dealing with boreal sediments) researchers. A part of the Volgian Stage should undoubtedly be included into the Cretaceous System, but the question is which one? Until the decision of the International Commission on Stratigraphy about the GSSP of the Berriasian Stage boundary, the ISC RF has no rights to solve this problem ignoring international agreements. This is a leitmotif of the paper. Consequently, the ISC RF decision of February 2, 1996, should be reconsidered, and the Volgian Stage should be returned into the general stratigraphic scale of Russia to be there until the Volgian–Tithonian and Berriasian–Boreal Berriasian zonal correlations will become cogent. Finally, I tried to substantiate the necessity of preservation of the Volgian Stage "status quo" not only in Russia. The Volgian Stage cannot be considered as a regional stage of Russia. The Volgian deposits cover territories of several countries, and an agreement between Russia, Canada, the United States, Denmark, Norway, and England is required on this issue. The ISC RF cannot solve the Volgian Stage problem ignoring other subjects of the international law.

Following J. Cope (1996), I suggest to introduce into the International stratigraphic scale the parallel stages in traditional nomenclature for the geological time intervals of a high biota differentiation. This idea seems sound, and I tried to substantiate this in the paper. Is the existence of national measures of length and weight more justified? They are just a tribute to tradition, but many countries do not hurry refuse them. The scale with parallel stages reflects a variety and unique character of past geological events. During million years, vast territories (blocks) of the Earth were much closer spaced, had a common geological history (tectonics, magmatism, sedimentation cycles, biota evolution, etc.) in distinction from other territories with their own history of geological development. The availability of scale with parallel stages, every of which has its own (original) essence, will enrich the historical geology. It should be emphasized that the parallel scales are free of terminological innovations. As a rule, names of parallel stages existed for a long time in "old" non-revised scales. To the contrary, new stage names are the result of the general scale revision that commenced at the end of the last century. I believe that most of stratigraphers realized during this revision that it was motivated by the national prestige rather than by striving to a universal convenience (Cope, 1996). The parallel scale will smooth the international faction. I consider it possible to borrow the way of official adoption of every parallel stage from initiators of the International stratigraphic scale revision, i.e., to connect it with the procedure of the GSSP selection.

Leaders of the International Commission on Stratigraphy made decision to finish the revision of the International stratigraphic scale by 2008. No one believes however that the revision will ever be finished. The general scale is inexhaustible and many generations of researchers will be busy with its perfection. This process will be determined by a progress in the "regional" stratigraphy. The general stratigraphic scale will be changed based on the regional scales. What if not this is demonstrated by experience of the last decade and by a destiny of some "Russian" stages of the Permian and Carboniferous systems? Thus, the further study of the Volgian Stage is necessary but not only desirable.

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Marine Paleogene Sediments from the Northern Slope of the Kuril Deep-Sea Basin, Sea of Okhotsk

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Abstract—Micropaleontology and lithology of rocks dredged near the Terpeniya Rise of the Kuril Basin slope are studied. Bio-, litho-, and seismostratigraphic records are used to distinguish the Paleocene–Lower Oligocene (Seismocomplex E), Upper Oligocene (Seismocomplex D), and Upper Oligocene–Middle Miocene (Seismocomplex C) sequences of the Lower Cenozoic sedimentary cover. The sequences are correlated with concurrent deposits of the region, and their paleotectonic settings are reconstructed. It is concluded that in the southern transitional zone of the Sea of Okhotsk there was a full succession of the Paleocene morphological structures, i.e., the shelf, slope, and deep-sea basin, which were separated from the ocean by the paleo-Kuril system of islands or submarine rises.

Key words: Paleogene, palynologic assemblages, diatoms, radiolarians, sediment accumulation, shelf, slope, deep sea basin.

INTRODUCTION

Geological and geophysical research of the Kuril Basin slope adjacent to Sakhalin, the Terpeniya Rise included, was intense in the 1970s, when the oil-and-gas prospecting works in the eastern Sakhalin shelf were in progress (Solov'ev *et al.*, 1974, 1975; Vasil'ev, 1981; Vasil'ev *et al.*, 1975; Ovcharenko *et al.*, 1975). It was found that the rise represents a horst-like uplift with a basement presumably composed of jasper, extrusive rocks, quartz–chlorite schists, hornfels, and granite ranging in age from the Paleozoic to Cenozoic. The basement is overlain by sedimentary cover that is subdivided into the lower sequence deformed into large gentle folds, and the upper one almost lacking deformations. The lower sequence is composed of tuffaceous argillites, sandstones, and siltstones bearing siliceous concretions. It is dated back to the early Oligocene–middle Miocene based on spores, pollen, diatoms, and malacofauna and correlated with the Arakai, Kholmsk, and Kurasi formations of Sakhalin. The upper sequence is composed of tuffaceous diatomites, tuffs, tuffaceous gravelstones, and breccia. Based on palynologic assemblages, this sequence is attributed to the Upper Miocene–Pliocene, thus being correlative with the Maruyama Formation. Using the generalized seismic profiles and dredging results, Kudel'kin *et al.* (1986, p. 10) suggested a model of Cenozoic sedimentation in the southeastern submarine margin of Sakhalin. They concluded that the earliest late Oligocene–early Miocene stage of Cenozoic sedimentation was related to "the initiation of a tephra-generating system in the Sea of Okhotsk plate and to the opening of the South Okhotsk (Kuril) deep-sea basin."

In this work, we generalized data of subsequent research carried out by the Pacific Institute of Oceanology, Far East Division, Russian Academy of Sciences (POI FED RAS). These are paleontological and lithologic characterization of samples from the Terpeniya Rise eastern slope, which were dredged during legs 37 and 39 of the r/v "Pervenets" in 1981–1982 (*Otchet o geologicheskikh...*, 1982; Bezverkhnii *et al.*, 1998; Tsoi *et al.*, 1998), interpretation of the regional CDP reflection profiling performed by "Dal'morneftegeofizika" in 1985, and deep-sea drilling data (*Ob"yasnitel'naya zapiska...*, 1994; Bakhanov *et al.*, 1990). The generalization results are used to refine and to revise in some cases the conclusions mentioned above.

CHARACTERIZATION OF THE STUDY REGION

The study region is situated in the junction zone of the Sakhalin tectonic system and the deep Kuril Basin (Fig. 1). The northern slope basement of the latter is composed of the Senonian (Bezverkhnii, 2000) volcano-plutonic formations of the Academii Nauk Rise that is a former island-arc system (Krasnyi, 1979; Zhuravlev, 1982; Bezverkhnii, 1997). On the slope oriented in a northeastern direction, there is the Terpeniya Rise, an escarpment of north-northwestern extension that represents a part of the East Sakhalin imbrication

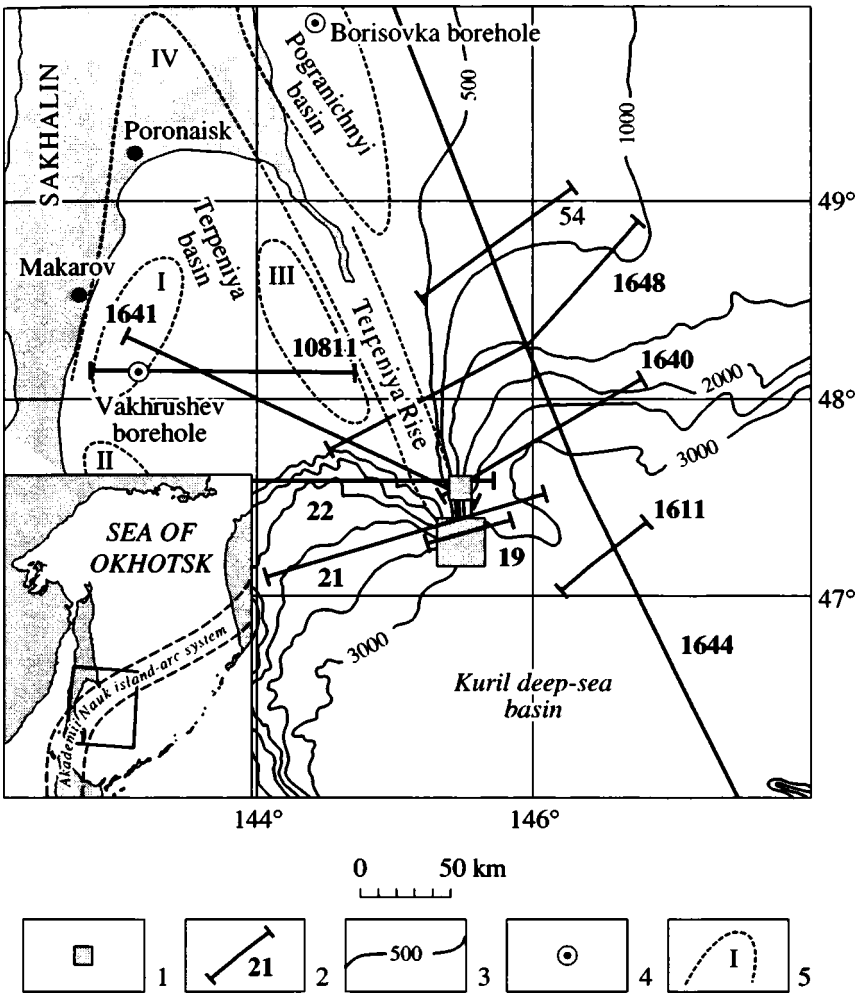


Fig. 1. Map of the study region: (1) dredging areas; (2) seismic profiles and their numbers; (3) isobaths; (4) boreholes; (5) troughs.

zone formed in the early Paleogene (Raznitsyn, 1982; Rikhter, 1986). The zone separates two large areas of shelf sedimentation: the eastern Pogranichnyi and the western Terpeniya basins, the latter corresponding to the Makarov (I), Starodubskoe (II), and Vladimirovo (III) troughs. The Terpeniya basin extends on land as the Poronaisk depression (IV). Dredge samples are from two areas of the study region with their center located at 47°11.3′ N and 145°28.0′ E. The escarpment slope is rather steep (up to 30°) and high (to 2600 m).

METHODS

The bio-, litho-, and seismostratigraphic data are generalized using the procedure developed in the POI FED RAS (Bezverkhnii *et al.*, 1980). The metric thickness of seismocomplexes on the shelf and slope were determined based on stratal velocities established by seismic refraction method (Popov, 1983; Bikkenina *et al.*, 1987). The distinguished seismostratigraphic units of the shelf, continental rise, seafloor, and slope of the Kuril Basin are correlated using a combination of seismic records, drilling and dredging data.

RESULTS AND DISCUSSION

The morphostructural basis of the Terpeniya Rise corresponds to a horst connected with the upthrow-thrust system of the Terpeniya Peninsula that is mainly composed of the Upper Cretaceous–Danian volcanogenic–sedimentary and extrusive rocks (Grannik, 1978). It is likely the Cretaceous rocks ($\bar{V}_P$ = 5.2 km/sec), which correspond to the rise acoustic basement. The basement surface is of a tectonic relief having the east-northeast trending structural plan. According to available seismic records, the basement is not exposed at the bottom of the study region (Figs. 2 and 3). The Paleozoic–Mesozoic intrusive rocks and products of contact metamorphism ($\bar{V}_P$ = 5.0–6.9 km/sec), samples of which have been dredged from the escarpment southern slope and considered previ-

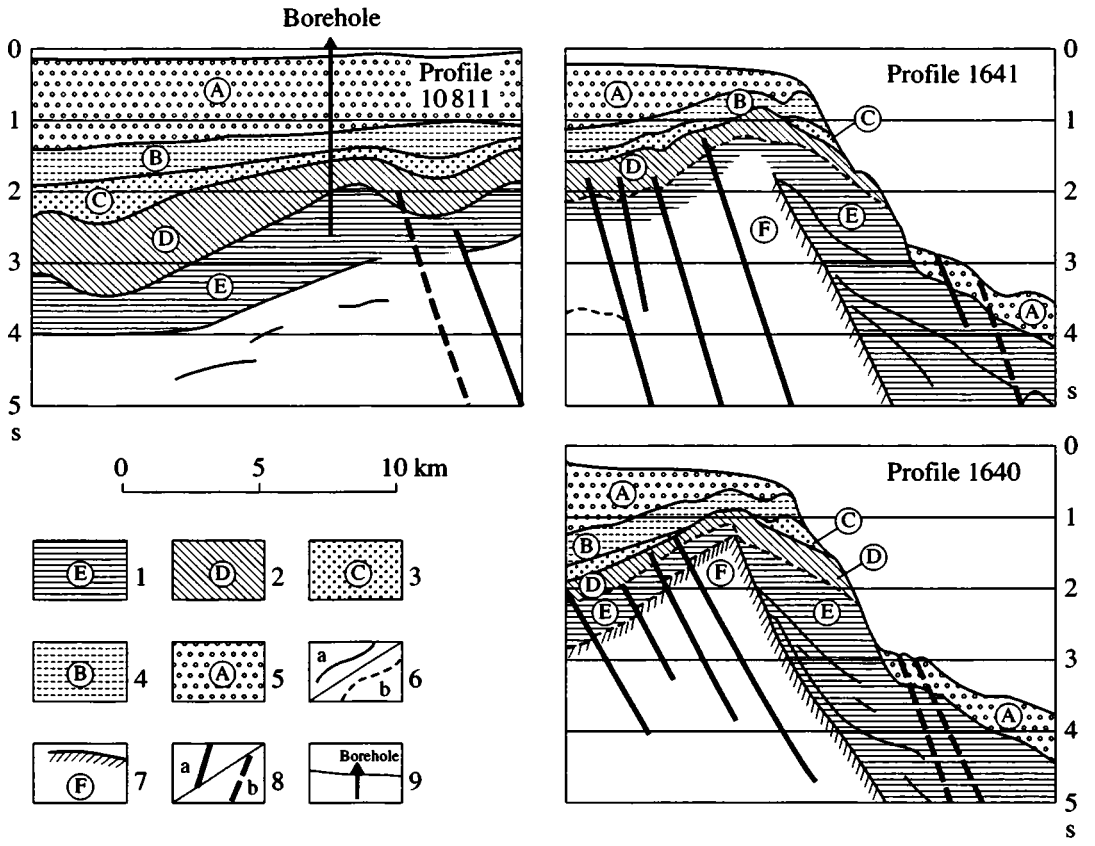


Fig. 2. Geologic interpretation of CDP seismic profiles 10811, 1641, and 1640 (fragments): (1–5) seismocomplexes; (6) seismocomplex boundaries reliable (a) and inferred (b); (7) index and surface of acoustically resistant sequence; (8) faults reliable (a) and inferred (b); (9) Vakhrushev borehole.

ously as index rocks of the latter (Vasil'ev *et al.*, 1975), were most likely transported by ice from the on-land structures around Sea of Okhotsk (Archikov and Stepanova, 1987).

The relict subaerial topography is recognizable in the basement surface of the northern horst area only (Solov'ev *et al.*, 1975). The horst was evidently formed in submarine conditions, and accumulation of sediments on its slopes terminated, when declivity of the latter exceeded the angle of repose of water-saturated sediments. In the adjacent shelf depressions, the surface of acoustically resistant sequence is commonly recognizable with a high degree of uncertainty, and in some cases, it is placed conventionally at the disappearance level of stable reflections.

The acoustically transparent sedimentary cover is composed of two sequences separated by a structural unconformity. The lower one is subdivided into seismocomplexes E and D, which reveal gentle synsedimentary folds with axes striking in a northwestern (to submeridional) direction. In the upper sequence, sedimentary strata are subhorizontal. The sequence is subdivided into seismocomplexes C, B, and A (Figs. 2–4).

THE PALEOCENE–LOWER OLIGOCENE (P_1–P_3^1) SEISMOCOMPLEX E

On the slope, Seismocomplex E is in the onlap relations with the surface of acoustically opaque sequence dipping at the angle of 30°. The medium- to high-amplitude reflectors in the complex are of a medium or short extension, grading into intermittent reflectors of medium amplitudes and complex shape eastward from the slope. In the lower part of the slope, the seismocomplex is up to 3.0 km thick ($\bar{V}_P$ = 4.0 km/sec).

The upper part of the sequence corresponding to Seismocomplex E is sampled by dredging at sites 2221–2227, 2230, and 2232 within the depth range from 2500 to 1350 m. The sampled argillites and siltstones bear both the detrital and authigenic grains of glauconite. Quartz, feldspar, muscovite, and ore minerals are abundant in argillites. Tourmaline, zircon, garnet, and fragments of carbonate, quartz–sericite, and quartz–feldspar rocks are scarce. Plagioclase is commonly zonal, of pyroclastic origin. Matrix of the rocks consists of clay minerals and is slightly calcareous. In some cases, cementing mass is represented by amorphous silica. The diatom internal molds are commonly

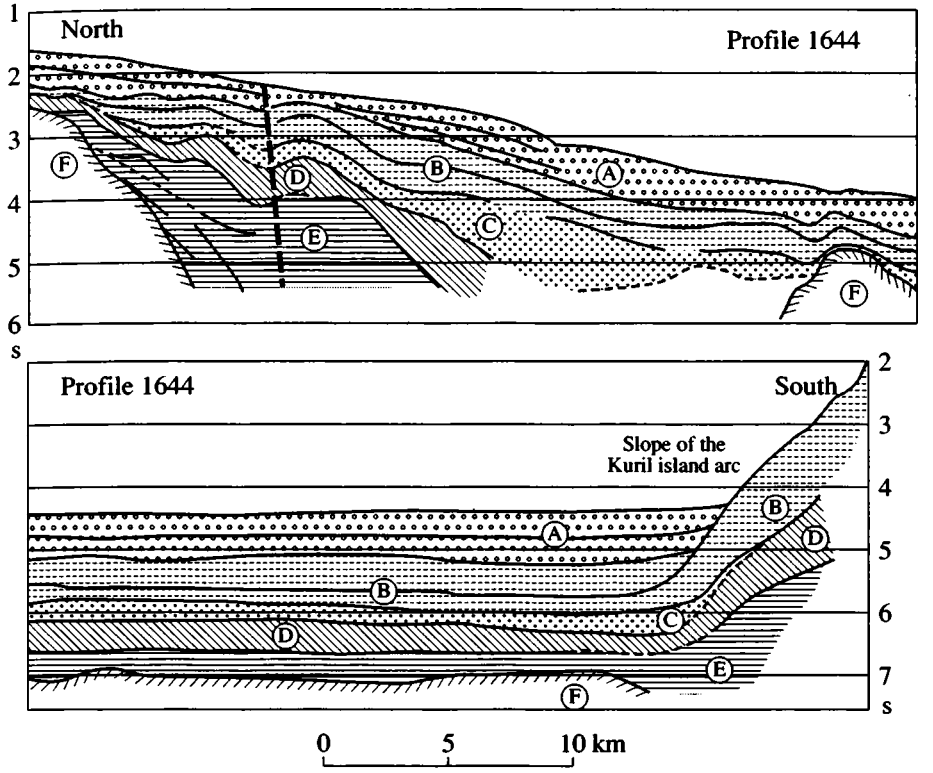


Fig. 3. Geologic interpretation of the CDP seismic profile 1644 (fragments; symbols as in Fig. 2).

filled with opal and carbonate. Argillites also bear tridymite–cristobalite (opal-C).

The *Triatriopollenites plicoides–Ulmoideipites krempii–Anacolosidites primigenius* palynological assemblage was recognized in six argillite and siltstone samples from Site 2227. Spores of form taxa *Triatriopollenites plicoides* Zakl., *T. plicatus* Krutsch, *T.* aff. *confusus* Zakl., *T. aroboratus* Pfl., and *T.* sp. is abundant and diverse among angiosperms of the assemblage and associate with less abundant *Tricolpites* sp., *Triporopollenites* sp., and *Tricolporopollenites* sp. The assemblage includes diverse flowering taxa of Ulmaceae (*Ulmoideipites krempii* Anders., *U. tricostatus* Anders., and *Ulmus* sp.), Juglandaceae (*Engelhardtia* sp., *Juglanspollenites* sp., *Caryapollenites* sp., *Platycaryapollenites* sp., and *Pterocaryapollenites* sp.), Betulae (*Betulapollenites* sp., *Alnuspollenites* sp., and *Coryluspollenites* sp.), Fagaceae (*Faguspollenites* sp. and *Castanopsis* sp.), and Ericaceae. The Olacaceae-like *Anacolosidites primigenius* Zakl. and *Anacolosidites* sp. are also characteristic taxa. Taxodiaceae (*Taxodiumpollenites* sp.) and Pinaceae (*Piceapollenites* sp., *Pinuspollenites* sp., *Cedripites* sp., *Abiespollenites*, and diverse *Tsugapollenites*) abundant among gymnosperms associate with *Ginkgocycadophytus* sp., *Podocarpidites* sp, and *Dacrydiumites* sp.

Sporogenous plants are mainly represented by taxa characteristic of the Cretaceous palynofloras: *Cyathidites minor* Coup., *C. australis* Coup., *Concavissimisporites asper* (Bolch.) Poc., *Laevigatosporites ovatus* Wills. et Webst., *L. ovoideus* Takah., smooth trilete *Leiotriletes* sp., and by Osmundaceae ferns (*Osmundacidites wellmanii* Coup. and *O.* sp.), club-mosses (*Leptolepidites verrucatus* Coup. and *Baculatisporites* sp.), and mosses (*Concavisporites* sp.).

The studied assemblage includes a considerable amount of Maastrichtian–Danian taxa among angiosperms and ferns, although typical Maastrichtian forms of "unica" and "oculata" groups, which still occur in small proportion in the Danian palynofloras, are absent. A significant abundance is also characteristic of *Anacolosidites*, the peculiar taxon dominating in the Eocene palynofloras.

The assemblage is close to the "Inter-conglomerate" palynoflora from the lower sandy siltstone member of the Poronaisk depression (*Tektonika i neftegazonosnost'...*, 1985) and to the assemblage from the lower and middle parts of the Lower Due Formation of the Uglegorsk section, Sakhalin (Brutman, 1976, 1986). Its characteristic feature consists in absence of form genera *Aquilapollenites, Mancicorpus*, and *Orbiculapollis* characteristic of the late Cretaceous palynofloras. Correlation of the *Triatriopollenites plicoides-Ulmoideipites krempii-Anacolosidites primigenius* assemblage with concurrent palynofloras from adjacent regions implies that the enclosing sediments are of the late

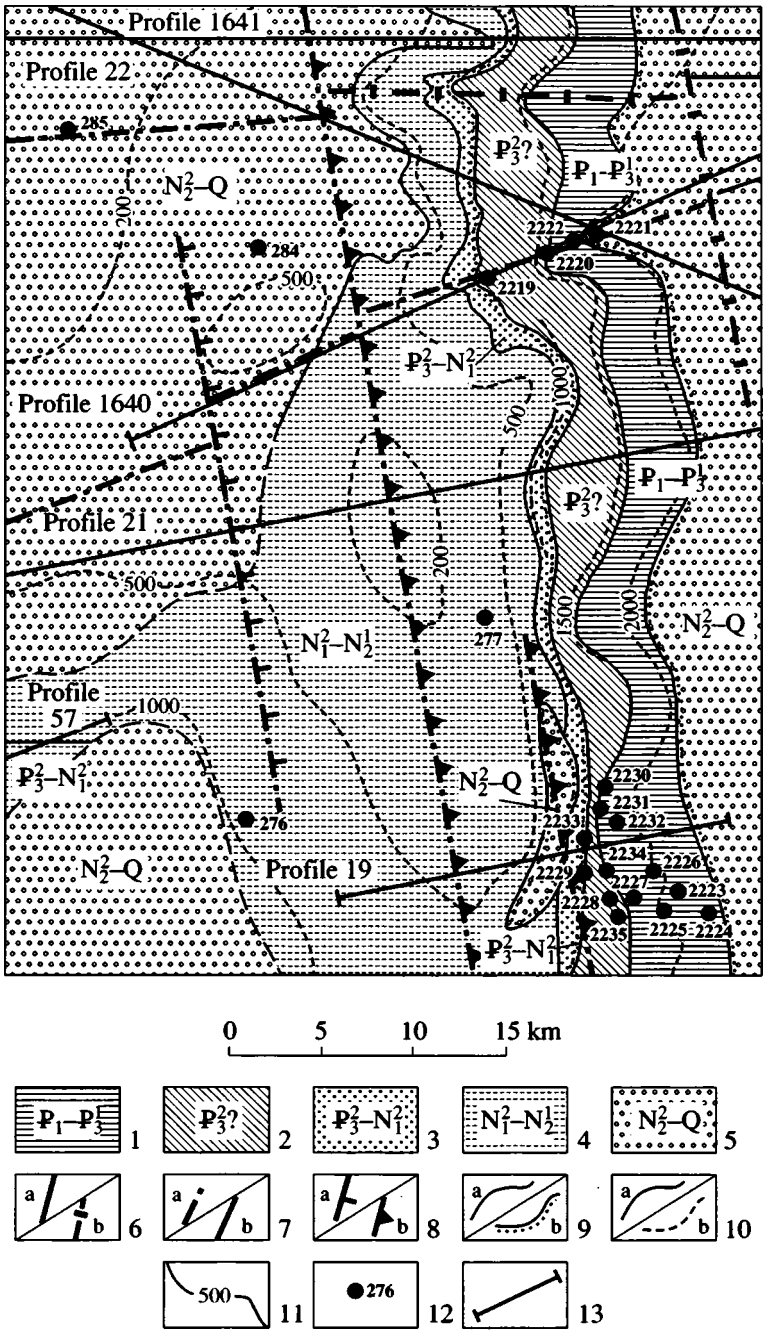


Fig. 4. Schematic geologic map of the southern Terpeniya Bank: (1) Paleocene–lower Oligocene, argillites, siltstones, tuffaceous diatomites; (2) upper Oligocene; (3) upper Oligocene–middle Miocene, tuffaceous diatomites, carbonate rocks; (4) middle Miocene–lower Pliocene, tuffaceous siltstones; (5) upper Pliocene–Quaternary, tuffaceous siltstones, tuffaceous diatomites; (6) faults traceable on the bottom reliable (a) and inferred (b); (7) faults below the bottom, reliable (a) and inferred (b); (8) normal (a) and upthrow (b) faults; (9) geologic boundaries conformable (a) and unconformable (b); (10) boundaries reliable (a) and inferred (b); (11) isobaths; (12) dredging sites; (13) seismic profiles.

Paleocene–early Eocene age. A considerable proportion of thermophilic taxa in the assemblage, e.g., of *Anacolosidites, Engelhardtiapollenites, Platycaryapollenites, Podocarpidites, Dacrydiumites*, and some other forms., is indicative of a moderately warm and humid climate. Abundant *Taxodium* specimens suggest presence of swampy plains nearby the sedimentation basin.

One argillite sample from Site 2227 yielded the taxonomically different *Picea gigantea–Pinus–Taxodium* palynologic assemblage. This assemblage is dominated by gymnosperms, mainly by Pinaceae, including abundant *Picea* sp. (*gigantea*), *Picea* sp., *Pinus* s/g *Haploxylon, P.* sect. *Strobus, Pinus* sp. (*minima*), *Pinus* sp., and diverse *Tsuga* forms: *Tsuga parva* Brutm, *T.*

sauerae Brutm, and *T. diversifolia* (Maxim.). *Taxodium* sp. is also abundant. The concomitant taxa include flowering plants *Betula* sp., *Alnus* sp., *Ulmus* sp., *Juglans* sp., *Carya* sp., *Tilia* sp., *Fagus* sp., and pollen of the form genus *Triatriopollenites*. Sporogenous plants are represented by a few Polypodiaceae and *Osmunda* sp.

According to taxonomic composition and abundant *Taxodium* forms, the assemblage is correlated with the Central Tamlevo Oligocene assemblage from the upper sandy siltstone member of the Poronaisk depression (*Tektonika i neftegazonosnost'...*, 1985) and with palynofloras from the lower part of the Arakai Formation of the Uglegorsk section, Sakhalin (Brutman, 1976, 1986), and from the lower part of the Gastello Formation, Makarov section, Sakhalin (*Opornyi razrez...*, 1992). The analogous assemblages are recorded in the upper part of the Amanina and lower part of the Gakkh formations (Tochilino and Kheislino palynocomplexes) in the western Kamchatka (*Atlas fauny...*, 1984; Brutman *et al.*, 1985). These features along with the lack of typical Paleocene–Eocene species in the considered assemblage suggest its early Oligocene age and moderately warm and humid climate during the host sediment accumulation.

Two radiolarian assemblages are established in samples from Site 2227. The first one from argillites and argillaceous siltstones includes abundant Spongodiscidae, while Cenosphaeridae, Lithelidae, and Stylodictyidae occur as single specimens. The second assemblage from argillites, siltstones, and phosphatic rocks consists of scarce specimens of all above-mentioned families associated with Heliodiscidae forms. The taxonomic composition of both assemblages indicates the sediment accumulation in a coastal sea zone with normal salinity (Bezverkhnii *et al.*, 1998). Marine diatoms from these samples are represented by half-dissolved valves of *Coscinodiscus, Stephanopyxis, Rhizosolenia*, and *Pseudotriceratium* genera (Tsoi *et al.*, 1998).

Seismocomplex E is correlated with the Paleocene–Lower Oligocene sandy siltstone sequence of the Poronaisk depression, which overlies with a structural unconformity the Mesozoic volcanogenic–sedimentary and volcanogenic sediments of the Ostrino and Khoi formations. The lower member of paralic deposits up to 0.9 km thick is composed here of interbedded sandstones, siltstones, and, rarely, carbonaceous argillites and coal seams. In these sediments, Brutman recognized the "Inter-conglomerate" palynoflora (*Tektonika i neftegazonosnost'...*, 1985) typical for the Paleocene *Alnipollenites–Anacolosidites* palynozone that corresponds to the lower and middle parts of the Lower Due Formation in the Uglegorsk region of Sakhalin, and has been assigned by Zaklinskaya to phases *a* and *b* of the Cenophytic stage III (*Razvitie flor...*, 1977).

The middle member up to 400 m thick conformably overlies the previous one. It is represented by siltstones intercalated with argillites and sandstones. The sandstones commonly yield glauconite that constitutes up to 30% of the rock volume; the siltstones and argillites contain inclusions of carbonaceous matter. In these sediments, Brutman (*Tektonika i neftegazonosnost'...*, 1985) recognized the Muz'ma palynoflora typical of the Eocene *Lygodium japoniciforme–Pinus* s/g *Haploxylon–Ulmoideipites* palynozone that includes palynologic assemblages from the upper part of the Lower Due Formation and from the Krasnopol'e and Takaradai formations of southwestern Sakhalin. The palynoflora corresponds to phases *c* and *d* of the Cenophytic stage III. The section crowned by the upper sandy siltstone member up to 410 m thick that is likely of the early Oligocene age has been accumulated in the littoral zone (*Tektonika i neftegazonosnost'...*, 1985).

In the Terpeniya Bay shelf, Seismocomplex E fills in and envelops forms of tectonically dissected topography at the surface of the acoustically opaque sequence, showing the onlap and, rarely, downlap relations with the latter. The sedimentary depressions outlined in structural maps of the seismocomplex are extending east to northeast in accordance with the basement structural plan (Solov'ev *et al.*, 1975). The thickness of Seismocomplex E increases west of the horst and reaches in places more than 4.0 km ($\bar{V}_P$ = 4.0 km/sec). In the outer shelf, the seismocomplex is disjunctively deformed and folded above the basement uplifts.

In the Vakhrushev borehole section, the eastern side of the Makarov Trough, the sequence of siliceous argillites, siltstones, and fine-grained sandstones is about 350 m thick and accumulated in a coastal area. This sequence bearing foraminifers and palynomorphs is correlative to Seismocomplex E. According to lithologic features, it is correlative with the Gastello (*Ob"yasnitel'naya zapiska...*, 1994) or Kholmsk and Nevel'sk formations (Bakhanov *et al.*, 1990), but its exact age is undetermined.

In the continental rise, Seismocomplex E is over 5.0 km thick ($\bar{V}_P$ = 4.0 km/sec) and fills a system of troughs termed by Kharakhinov and others (Baboshina *et al.*, 1985) as a "graben of continental slope". Seismic records of the complex are strongly complicated here, suggesting presence of diffracted waves and the reflection correlation loss. These features are most likely associated with the recent slumping and faulting.

On the basin floor, Seismocomplex E is up to 1.1 km thick ($\bar{V}_P$ = 4.3 km/sec), showing the extended parallel reflections mainly of medium to short amplitude. On the floor uplifts, it overlies the acoustically opaque sequence with the onlap or scarce offlap relations. Within the abyssal plain, an enhanced seismic transparency of the seismocomplex is increasing toward its base. Seismic patterns, which are analogous to those in the basal, acoustically transparent part of sedimentary cover in the northwestern Pacific, suggest the accumulation of seismocomplex in a deep-sea environment far from provenances of terrigenous material.

Near the island base, the thickness of Seismocomplex E decreases to 500 m. Below the Greater Kuril Arc, the basement surface is plunging, whereas reflectors near the seismocomplex top sharply rise, and the total unit thickness grows to the values over 2.3 km ($\bar{V}_P$ = 4.7 km/sec).

THE UPPER OLIGOCENE (P_3^2?) SEISMOCOMPLEX D

On the slope, Seismocomplex D about 500 m thick ($\bar{V}_P$ = 2.3 km/sec) reveals mainly intermittent medium-amplitude reflections. Its contacts with the over- and underlying complexes are unclear. It transgressively overlies the surface of the acoustically opaque sequence, suggesting the onlap relation. Erosional boundaries in the seismocomplex have not been determined on the slope, where the sequence corresponding to Seismocomplex D has not been sampled. Having no seismograms linking the shelf and adjacent land sections, we defined the complex relationships with the on-land formations based on biostratigraphic correlation of the over- and underlying sequences.

On the adjacent island territory, Seismocomplex D is likely correlative with the Upper Oligocene sediments: the siliceous argillite sequence of the Poronaisk depression, which is about 0.7 km thick and crowned by a member of tuffaceous siltstones and andesite–dacite tuffs (*Tektonika i neftegazonosnost'...*, 1985); the Pilenga Formation of bathyal deposits up to 1000 m thick in the on-land part of the Pogranichnyi trough, which are represented by argillites and siltstones bearing carbonate and lens-like quartz–chalcedony concretions and intercalated with tuff beds in the upper part. The formation is dated back to the late Oligocene. The presence of the early Oligocene sediments in its lower part is assumed (Gladenkov and Gladenkov, 1999). In the Makarov region, Seismocomplex D is correlative with the marine Upper Oligocene–Lower Miocene sediments of the Sergeevo Group, the Kholmsk, Nevel'sk, and Chekhov formations included. Volcanogenic–sedimentary deposits of the Chekhov Formation of a high lithologic variability are impregnated by sills, intrusive and subvolcanic bodies of gabbro–diorite, diorite, andesite, and basaltic andesite compositions (*Ob"yasnitel'naya zapiska...*, 1994).

In the shelf, reflectors in Seismocomplex D are mainly of a high amplitude, extended, and not frequent. The complex is up to 1.2 km thick here ($\bar{V}_P$ = 2.3 km/sec). It conformably overlies Seismocomplex E or is in onlap relations with the acoustically opaque sequence. On slopes of the basement uplifts, the seismocomplex reveals chaotic reflectors of a complex configuration, which likely suggest the sharp lithofacies variability of corresponding volcanogenic–sedimentary deposits and synsedimentary folding and faulting. Within the positive morphostructures and in uplifted side parts of troughs, there are local toplap seismic patterns most likely associated with the subaerial erosion. This is evident from truncation of oblique in-phase axes of seismic waves by overlying subhorizontal reflections of the seismocomplex.

In the Vakhrushev borehole section, Seismocomplex D corresponds to the sequence of tuffs, tuffaceous sandstones and agglomerates, opokas, basic extrusive rocks, and tuffaceous breccias. The sequence is 820 m thick and unconformably rests upon the underlying rocks. It is correlated with the Kholmsk (*Ob"yasnitel'naya zapiska...*, 1994) and Chekhov (Bakhanov *et al.*, 1990) formations, although its age is undetermined.

In the continental rise, Seismocomplex D is up to 1.6 km thick ($\bar{V}_P$ = 2.3 km/sec), dashed by medium- to high-amplitude intermittent reflections with widespread diffracted waves. In the basin, it is up to 0.8 km thick ($\bar{V}_P$ = 3.0 km/sec), showing the extended parallel reflections of medium amplitudes. On the Greater Kuril Arc slope, Seismocomplex D conformably rests on the underlying sequence and transgressively overlies the acoustic basement surface forming an inverse clinoform up to 2.8 km thick ($\bar{V}_P$ = 3.2 km/sec).

The upper part of acoustically transparent sedimentary cover that includes seismocomplexes A, B, and C, is subhorizontal. It envelops gentle uplifts and fills in the depressions in underlying surface. This part of sedimentary cover is almost free of deformations.

THE UPPER OLIGOCENE–MIDDLE MIOCENE (P_3^2–N_1^2) SEISMOCOMPLEX C

On the Sakhalin slope, Seismocomplex C with medium-frequency reflectors is 400 m thick ($\bar{V}_P$ = 2.0 km/sec). Its lower part sampled at sites 2228, 2229, and 2233 within the depth range from 1250 to 1000 m represents a sequence of tuffaceous or pure diatomites presumably resting on a basal conglomerate bed. The clastic material composition is analogous in the argillite–siltstone and tuffaceous diatomite sequences. In both, matrix is composed of altered volcanic material in tuffaceous diatomites and of diverse clastic material in argillites. Pebbles and boulders of conglomerate correspond in composition to gabbro–diabase, diorite, diabase porphyrite, granite, granite–porphyry, basalts, acidic extrusive rocks, and siltstone hornfels. The matrix composition in conglomerate corresponds to sandstone or sandy tuffaceous diatomite. Palynomorphs have not been encountered in the lower part of the tuffaceous diatomite sequence. Diatomites and tuffaceous diatomites dredged at sites 2228 and 2229 yield the diatom and radiolarian assemblages.

I.B. Tsoi correlated the diatom assemblage with that of the Upper Oligocene–Lower Miocene *Thalassiosira prefraga* Zone, which is 24.0–20.3 Ma old. A correlative assemblage is known from the lower part of the

Bora Formation of the Pogranichnyi trough (Gladenkov and Gladenkov, 1999). According to their composition, host sediments accumulated in the sea basin with normal salinity. The predominance of planktonic neritic species constituting from 66 to 73% of the assemblage suggests a habitat in the outer shelf of an open sea (Tsoi *et al.*, 1998). The same samples yield radiolarians of the *Lipmanella japonica conica–Gondwanaria dogieli* assemblage of the Oligocene–early Miocene. V.V. Shastina noted among them the redeposited Eocene species and concluded that conditions were approaching the oceanic environment (Bezverkhnii *et al.*, 1998).

The seismocomplex upper part sampled at sites 2229 and 2235 within the depth range from 1250 to 850 m is composed of tuffaceous diatomites and calcareous rocks. The sediments contain bivalve mollusk remains and two diatom assemblages of the *Crucidenticula kanaya* and *Denticulopsis prelauta* zones. The associated radiolarian assemblage is correlative with that of the *Calocycletta costata* Zone, and palynomorphs represent the *Picea–Fagus–Ulmus*–Juglandaceae and *Fagus–Pinus*–Taxodiaceae palynofloras. Mollusks and diatoms are characteristic of a relatively deep-water environment and fairly high temperatures of surface water (Tsoi *et al.*, 1998). The latter palynoflora is typical of the climatic optimum at the early–middle Miocene boundary.

In the central troughs areas of the Terpeniya Bay shelf, Seismocomplex C is up to 400 m thick ($\bar{V}_P$ = 2.0 km/sec), conformably resting on Seismocomplex D. In their flanks, complex C shows the toplap relations with older deposits. The low-amplitude medium-frequency reflectors in the seismocomplex were observed in the central areas, and parallel reflectors of medium- to high-amplitude are characteristic of the rise flanks.

According to data on the Borisovka borehole, an upper part of the seismocomplex is represented in the shelf area of the Pogranichnyi trough by a sequence about 800 m thick that is composed of siltstones, argillites, and sandstones bearing marine mollusks, foraminifers and the Langry palynologic assemblage (Brutman, 1986; *Opornyi razrez...*, 1992). Paleontological materials suggest the early–middle Miocene age of enclosing deposits. In the Vakhrushev borehole section (Terpeniya Bay), the sequence corresponding to Seismocomplex C overlies with a hiatus sediments of the Sergeevo Series (Bakhanov *et al.*, 1990).

In the continental rise, Seismocomplex C is universally distributed. Complicated seismic patterns with widespread diffracted waves are recorded for its lower part, and parallel high-amplitude reflectors are characteristic of the upper one. The complex up to 1.2 km thick ($\bar{V}_P$ = 2.0 km/sec) conformably overlies the older deposits. In the deep-sea basin, the thickness and seismofacies of the complex are analogous to those of the underlying unit. In the rise of the Greater Kuril Arc, Seismocomplex C forms a simple clinoform up to 300 m thick ($\bar{V}_P$ = 2.0 km/sec). It regressively overlies the lower complex, as one can judge from the onlap patterns.

The overlying seismocomplexes and corresponding sequences (table) have been characterized in a previous publication (Bezverkhnii *et al.*, 1998).

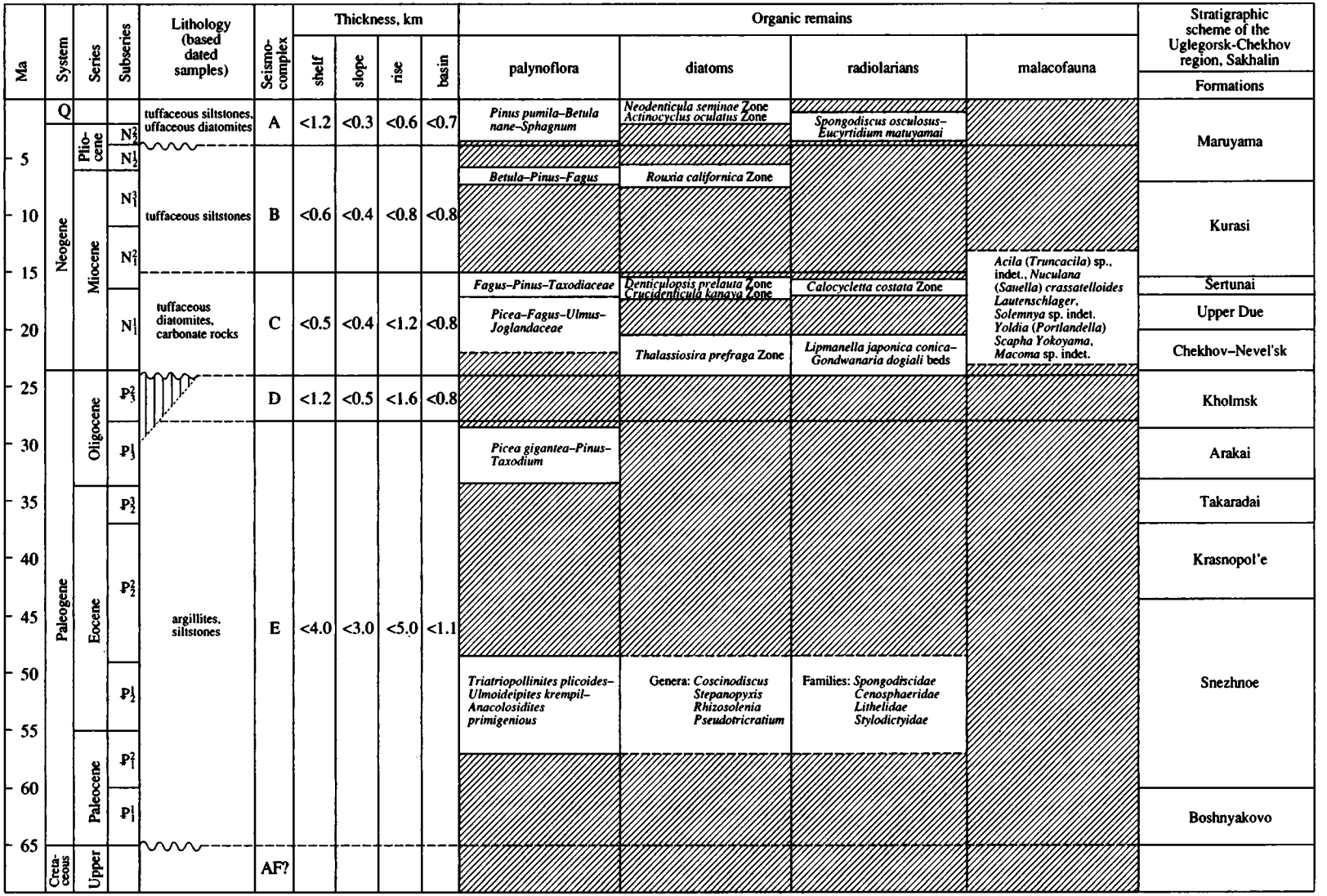
DISCUSSION

The structural unity of the Terpeniya Rise and East Sakhalin tectonic imbrication zone is beyond question. At the Cretaceous–Paleogene boundary time, the oceanic, island-arc, and marginal-sea structures of different age experienced faulting and thrusting. The hypabyssal gabbro–granite intrusions emplaced at that time as well gave rise to a partial consolidation of the zone and to formation of the acoustically opaque sequence. The consolidation was most intense in the northern areas, where a mountainous land (pre-East Sakhalin mountains) was formed. The amplitude of uplift decreased southeastward, and the structure most likely did not emerge above the sea level in the south. The structural plan of sedimentary basin has changed: the slope of the Akademii Nauk island-arc system of the Senonian time that is extended in the northeastern direction was divided at that time by the transverse upthrow-thrust structure of the present-day Terpeniya Rise. The sediment accumulation terminated within the northern Terpeniya Rise owing to its elevation in the latest Oldonian time of the Danian (Grannik, 1978). Sediments could not be deposited on steep submarine slopes of the rise as well and slid down to its foot. The sedimentary sequence that accumulated here overlapped the steeply rising surface of the acoustically opaque sequence.

In adjacent shelves, tectonic thrusting and granitization were most likely much less intense. The basement surface is reliably detectable here only in the positive morphostructures. The comparison of the parametric Magadan borehole section in the southern flank of the North Okhotsk trough (Bol'shakov *et al.*, 1989) with the corresponding seismogram shows that the Cretaceous sedimentary units in the trough depocenter are acoustically transparent. Thus, the gradual decrease of reflectors frequency downward in the Terpeniya Bay shelf depressions most likely indicates that the Cretaceous–Cenozoic sequence is continuous.

On the studied slope, the lower part of the acoustically transparent sequence is represented by Seismocomplex E that was sampled in its upper part only. The corresponding stratigraphic unit accumulated during the late Paleocene–early Oligocene in a coastal sea zone with normal salinity near a mountainous land composed of volcanogenic–sedimentary formations, sediments of which bear clastic material derived from the Upper Cretaceous Rymnik, Berezovka, and Kotikovo groups of the East Sakhalin mountains. Since the

Stratigraphic scheme of Cenozoic deposits in the northern slope of the Kuril Basin



Note: Hatched intervals are lacking paleontological records.

complex was sampled in its upper part only, it likely includes the Lower Paleogene deposits, as we believe. The transgressive onlap relations between the complex and acoustic basement surface in the rise (Fig. 2) and basin (Fig. 3) slopes indicate that sediments accumulated before the rise formation, i.e., in the pre-Cenozoic time.

The thickness variations of Seismocomplex E in the composite profile crossing the Poronaisk depression and Kuril Islands, relationships of the complex with acoustically resistant sequence, and seismofacies zoning enable a reliable paleotectonic reconstruction. According to such a reconstruction, a full set of marginal sea morphostructures, i.e., the coastland, shelf, continental slope and rise, and deep sea basin, which were separated from the Pacific by the pre-Kuril island or submarine arc, existed in the studied region during the Paleocene.

The upper part of the deformed sedimentary cover is represented by Seismocomplex D, characteristic of which are complicated wave interference patterns. The erosional surface at the top of corresponding sequence is commonly recognizable within positive morphostructures of the shelf. Factors responsible for that erosion event were likely the sea level drop in the late Oligocene and tectonic movements, which produced structural unconformity in the Terpeniya Bay shelf. The basement uplifting beneath the Terpeniya Rise, which was still active in the Neogene, is likely a consequence of thrusting along the fault system of northwestern strike.

Seismocomplex D of a regional distribution is correlated with a greater part of the Sergeevo Group sequence that includes the Arakai, Kholmsk, Nevel'sk, and Chekhov formations. The upper part of the group is of a problematic age, because of the Paleogene–Neogene boundary position is undetermined in Sakhalin. According to decisions of the interdepartmental stratigraphic conferences, this boundary was placed at the Kholmsk Horizon base in 1974 and at the top of this subdivision in 1998. There are opinions that the Nevel'sk and lower portion of the Chekhov formations in the upper part of the Sergeevo Group are of the Paleogene age. For instance, Serova (1978) noted significant taxonomic changes among mollusks and foraminifers crossing the boundary between the Nevel'sk and Chekhov regional stages and considered this level as probably corresponding to the Paleogene–Neogene boundary. Krasilov *et al.* (1984) described a macrofossil assemblage from the top of the Chekhov Formation in the Makarov region and defined its age as corresponding to the Oligocene Stage or Oligocene–Miocene boundary. Analyzing dinoflagellates from the Kholmsk and Nevel'sk formations of the same region, Kurita *et al.* (2000) concluded that they originated in the mid-late Oligocene 27.0–25.4 Ma ago. Relying on radiologic estimates they assign the Chekhov Formation that is, in opinion of many researchers, a volcanogenic analogue of the Nevel'sk Formation to the late Oligocene–early Miocene.

Our results additionally support the last viewpoint. According to data described above, the rock sequence of Seismocomplex D is of late Oligocene age on the Terpeniya Rise, because it overlies the siltstone–argillite sequence of the Paleocene–lower Oligocene (Seismocomplex E) and is below the tuffaceous diatomites (Seismocomplex C) bearing diatom assemblages of the *Thalassiosira prefraga* Zone. The latter spans the time interval of 24–20.3 Ma (Akiba, 1986) and corresponds to the terminal late Oligocene–early Miocene (Tsoi *et al.*, 1998). If the seismostratigraphic correlation between the Vakhrushev and Terpeniya sections is correct, then tuffaceous sediments of the Chekhov–Nevel'sk sequence correspond to Seismocomplex D of the upper Oligocene. In the shelf, the latter reveals distinct signs of subaerial erosion and the Oligocene–Lower Miocene radiolarian assemblage from the overlying sequence sampled on slopes includes the redeposited Eocene taxa. Therefore, V.O. Savitskii was likely correct in his assumption (oral communication of 1986) that the regional erosional surface at the Sergeevo Group top was formed during the period of a considerable sea-level drop 30–25 Ma ago (Vail *et al.*, 1977). However, the ultimate age of the boundary in question, which is of great practical importance, can be determined after additional research, e.g., after the stepwise dredging of the unique Cenozoic section on the eastern slope of the Terpeniya Rise.

CONCLUSION

Despite a somewhat conventional dating of the distinguished sequences (table), because they are sampled fragmentarily, the considered data elucidate conditions of the Early Cenozoic sedimentation on the northern slope of the Kuril Basin adjacent to Sakhalin.

The structural plan of the study region was mainly formed during the Early Paleogene tectonic movements, which created the East Sakhalin thrust–fault zone of a northwestern strike, which grades in the south into the Terpeniya Rise. Imbricated here are the Upper Cretaceous volcanogenic–sedimentary deposits of the ancient Sea of Okhotsk that represented a sea basin, the floor of which was dissected by a system of distinct depressions and uplifts extending northeast. The southernmost ridge, a remnant of which is the Akademii Nauk Rise, separated the basin from the ocean. The Cenozoic sedimentation under conditions of normal salinity was in progress on the outer shelf and slope above the Terpeniya Rise basement and in the continental rise and deep-sea basin that was inherited from the Upper Cretaceous sea but already separated from the ocean by the pre-Kuril Arc. The general trend of geologic evolution was persistent throughout the Cenozoic. The on-land part of the imbrication zone was a provenance of terrigenous material and an archipelago of volcanic islands supplied a significant amount of pyro-

clastic material. The contribution of ice-rafted clastic material was most likely significant since the late Oligocene climatic pessimum. The greatest part of the slope did not rise above the sea level in the Cenozoic. Sedimentation on the positive morphostructures of the shelf was repeatedly interrupted. The late Oligocene, erosion was most likely associated with the eustatic regression and tectonic movements responsible for origin of the main structural unconformity in the studied rock sequence.

The problems discussed in this work are far as yet from the exhaustive solution. We are going to continue geological and geophysical investigations of the Terpeniya Rise sedimentary cover in order to understand in detail the Cenozoic evolution of the Sea of Okhotsk region.

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Key Reflecting Horizons in Sedimentary Fill Seismic Records of the Black Sea Basin (Correlation and Stratigraphic Position)

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Abstract—The seismic CMP is the main method of the geological sections study in the Black Sea, because drilling data on the deep basin part are scarce. Accordingly, stratigraphic subdivisions of sedimentary cover can be defined and correlated here based on key reflecting horizons. In the seismic record of Cenozoic sediments, four key reflecting horizons are distinguished at the base of the Quaternary (C), at the base of the upper Miocene, (I) at the top of the lower Miocene (Ia), and at the top of the Eocene (IIa). The reflecting horizon at the base of Cenozoic sequence (10–12 km thick) corresponds to the top of the Cretaceous overlain by Paleogene sediments (reference reflecting horizon III) in some areas or represents the eroded Mesozoic surface in the others (unconformity U). The key reflecting horizons divide sedimentary fill of the basin into five large complexes of the Quaternary, upper Miocene–Pliocene, middle Miocene, Oligocene–lower Miocene (Maikop Group), and Paleocene–Eocene. Data enabling a more detailed stratigraphic subdivision of sedimentary sequence are unavailable. Mostly simple, almost horizontal attitude of beds well traceable throughout the basin is favorable for correlation of reflecting horizons. All the defined key reflecting horizons and their lateral changes, correlation with onshore and offshore boreholes sections, and some peculiarities in bed attitude are described in detail. It is established that the thickest and most complete Cenozoic section is characteristic of the sea abyssal part, where key reflecting horizons are persistent and perfectly traceable along the strike. After additional study, they can probably be traced to the continental slope and even to the shelf.

Key words: Black Sea, seismic reflecting horizons, Paleocene–Eocene, Maikop Group, Miocene, Pliocene, Quaternary.

INTRODUCTION

In distinction from surrounding land, geological structure of the Black Sea basin can be studied based, almost exclusively, on the analysis of seismic sections. Bottom tectonics and sedimentary fill stratigraphy of the Black Sea became clear only at the beginning of 1980s, when systemic CMP sounding over the entire basin yielded seismic profiles of sedimentary cover structure to depths of 15 km (Fig. 1). The results of works have been already discussed (Tugolesov *et al.*, 1985; Gorshkov *et al.*, 1989). The seismic survey revealed all the tectonic structures of the second and third orders: two main depocenters of the Black Sea, which intensely subsided since the beginning of the Paleogene, several troughs formed in the Oligocene and intervening swells, numerous smaller uplifts, linear folds within troughs, and mud volcanoes.

After the first comprehensive analysis of seismic records (Tugolesov *et al.*, 1985), there was a series of subsequent works, which do not change, however, the former inferences. The works were mostly dedicated to inferable geodynamic models, discussion of which is beyond the scope of this work. Additional seismic records obtained during the last two decades specified only some details of the basin structure. Simultaneously, they showed that the previous analysis should be supplemented with data on correlation and stratigraphic position of defined reflecting horizons. To fill this gap is the purpose of the paper.

KEY REFLECTING HORIZONS

In seismic records of Cenozoic sediments, there are distinguishable four key reflecting horizons: at the base of the Quaternary (C), at the base of the upper Miocene (I), at the top of the Oligocene–lower Miocene sequence referred to the Maikop Group (Ia), and at the top of the Eocene (IIa). The base of the Cenozoic sequence is marked almost everywhere by the reflecting horizon, which corresponds locally to the top of the Cretaceous sequence (reflecting horizon III) conformably overlain by Paleogene sediments and usually represents the eroded Mesozoic surface (reflecting horizon U).

The reference reflectors divide the Cenozoic sedimentary sequence of the basin into five large complexes of the Quaternary (1), upper Miocene–Pliocene (2), middle Miocene (3), Oligocene–lower Miocene or the Maikop Group (4), and Paleocene–Eocene (5).

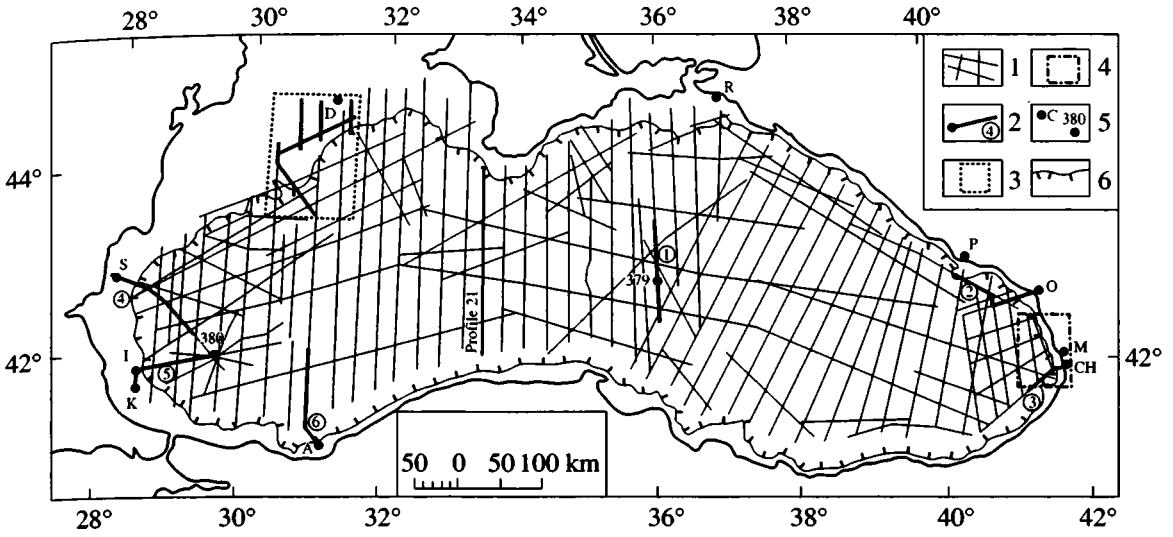

Fig. 1. System of seismic CMP profiles carried out by the State Scientific Center *Yuzhmorgeologiya*: (1) system of seismic profiles; (2) seismic profiles, schematic time sections of which are shown in Figs. 4 and 5 (sections numbers are encircled); (3) Danube delta and fan, where time sections of seismic profiles shown in Fig. 6 are correlated; (4) the eastern Black Sea area shown in Fig. 11; (5) boreholes used for correlation of key reflecting horizons: (A) Akchakhodzha-1, (D) Desantnaya, (I) Igneada, (K) Karadeniz, (M) Maltakva, (O) Ochamchire-4, (P) Pitsunda-2, (r) Rifovaya, (S) Samotino-more, (CH) Choloki, (379, 380) DSDP holes; (6) shelf break.

Data enabling a more detailed stratigraphic subdivision of the sequence are unavailable. The age of beds distinguishable in seismic time sections can be determined only by means of their correlation with known borehole sections. Three DSDP holes drilled in the Black Sea penetrated a relatively thin upper part of the sedimentary sequence, and fossils recovered are inappropriate for reliable and accurate age determination of boundaries, the base of the Quaternary included. Thus, only the correlation with onshore and shelf boreholes sections can settle the problem. The procedure is difficult, however, because the Cenozoic sediments thickness in the deep basin is incomparably greater than in the shelf and surrounding land, and many units frequently and completely pinch out at basin slopes. Nevertheless, some key reflecting horizons are locally traceable from onshore and shelf boreholes to the abyssal part of the sea.

Factors favorable for that correlation are the undisturbed, almost horizontal attitude of beds and considerable lateral extension of seismic reflecting horizons in the greater part of the basin. In well-studied sections of on-land Cenozoic troughs bordering the Caucasus, lithology of the defined sedimentary formations is laterally persistent. Some units of a small thickness, for instance, the carbonate members or even beds, are frequently traceable here from one section to another within the thick low-calcareous sandy-clayey Chokrakian–Karaganian sequence of the Terek oil-bearing region. In opinion of B.P. Zhizhchenko, these beds are strictly confined to particular stratigraphic levels. He also established that the 10- to 20-m-thick Alkun marly-dolomitic member at the base of the middle unit of the carbonate-free Maikop Group is traceable from the Caspian Sea to the Belaya River in the Kuban region, i.e. over a distance of hundreds of kilometers (Zhizhchenko, 1969).

The inferable sedimentation conditions in the Terek and West Kuban troughs are similar to those of the Black Sea basin despite the greater depths and permanent subsidence in the latter. Consequently, it is possible to expect that sedimentary formations and separate beds are of a more persistent lithology and better traceable in the Black Sea basin.

Indeed, the seismic sections of the Black Sea basin depict numerous reflectors traceable over large distances. Some of them, most distinct and continuous, are defined as key reflecting horizons. Despite the irregular dynamic acutance of reflectors over the seafloor area, the horizons are traceable also in areas of vague seismic records owing to a thick grid of profiles. Naturally, they are correlated with boundaries in borehole sections with a different precision. It is necessary to take into account therewith a possible error in assessment of the reflecting horizon depth, when calculating the latter based on seismic wave velocities in the sedimentary sequence. The statistical uncertainty of 15 m for the depth of 1 km below seafloor multiply increases for depths of 10 to 12 km.

The description of seismic units is accompanied in this work by schematic but not original seismic profiles, because the journal format is inappropriate for presentation of long profiles with original seismic records and vertical scale insignificantly exaggerated relative the horizontal one, which were factually used for correlation and stratigraphic positioning of seismic reflecting

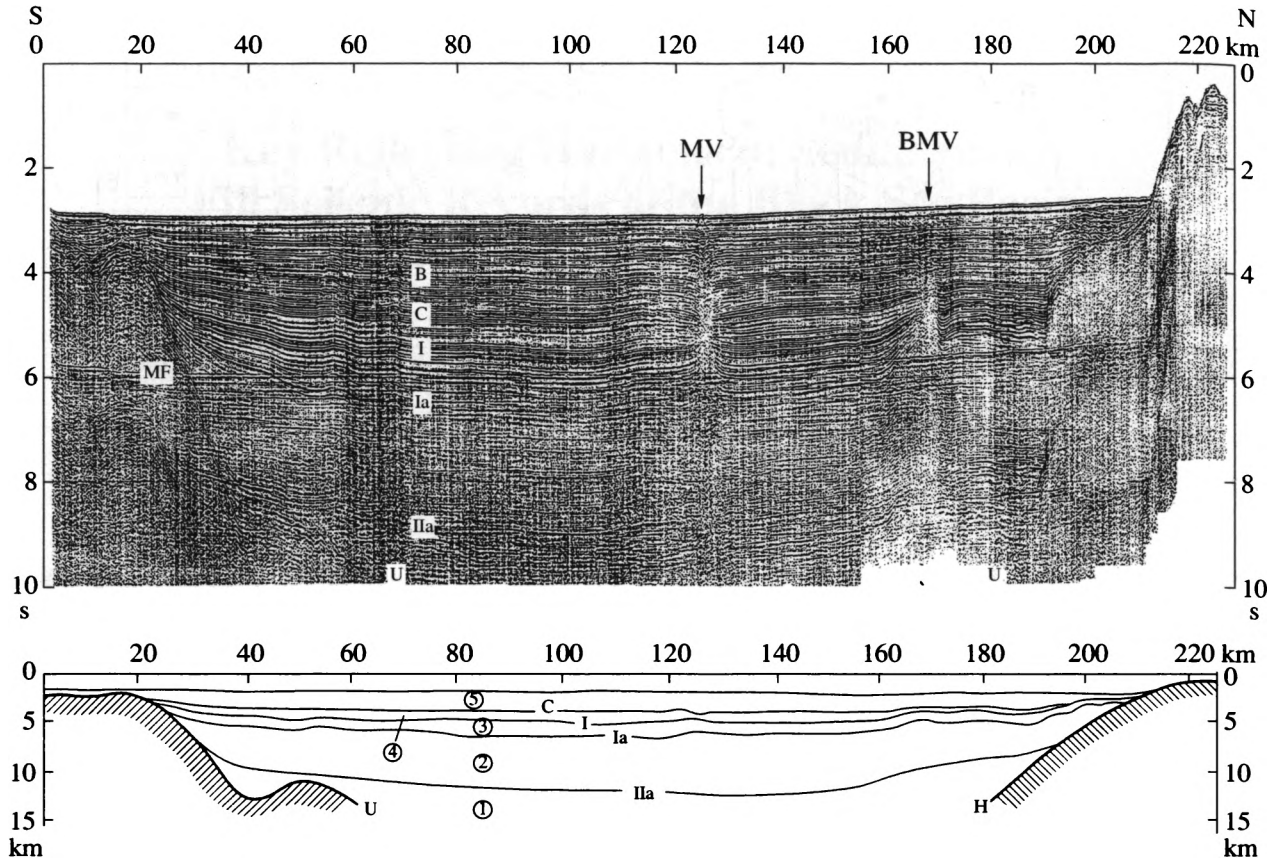

Fig. 2. Time section of CMP seismic profile 579 21 across the West Black Sea Basin (for location see Figs. 1, 3, 7–10); key reflecting horizons: (B) within the Quaternary sediments, (C) the base of the Quaternary, (I) the base of the upper Miocene, (Ia) the surface of the lower Miocene (top of the Maikop Group), (IIa) Eocene surface, (U) Mesozoic surface, (MF) upper boundary of multiple reflections ("multiple floor"); (MV) mud volcano; (BMV) buried mud volcano. The vertical scale is two times exaggerated relative the horizontal one. Encircled numbers correspond to sedimentary complexes: (1) Paleocene–Eocene, (2) Oligocene–lower Miocene (Maikop Group), (3) middle Miocene, (4) upper Miocene–Pliocene, (5) Quaternary.

horizons. One condensed and strongly reduced seismic section is presented in Fig. 2 to demonstrate the character of seismic records obtained for sedimentary cover of the Black Sea basin.

KEY REFLECTING HORIZONS III (THE TOP OF THE CRETACEOUS) AND U (ERODED SURFACE OF MESOZOIC DEPOSITS)

The reflecting horizon III corresponding to the top of the Cretaceous sediments is registered by many seismic profiles in the shelf and upper part of the continental slope of the Black Sea near Georgia. Within limits of the Ochamchire arch shelf, it is virtually horizontal like all overlying reflecting horizons. The onshore Borehole Ochamchire-4 penetrated the top of Cretaceous deposits at depth of 1900 m. This borehole also penetrated all other stratigraphic boundaries correlative with key reflecting horizons in the abyssal part of the basin: the top of the Eocene at depth of 1780 m, the top of the Maikop Group at 1350 m, the eroded roof of the Sarmatian Stage at 680 m, and the base of the Quaternary at 250 m. Marine seismic profiles begin 8–10 km away from the shore in this area. However, horizontal reflecting horizons continued to the borehole section coincide, within the seismic resolution limits, with above boundaries (Fig. 4, section 2).

Reflecting horizon III (like overlying ones) can be reliably traced owing to three intercrossed CMP profiles to the Gudauta arch. At first approximation, seismic records can be correlated here with the section of the coastal Borehole Pitsunda-2 via seismic CMP profiles obtained in the 1970s (Figs. 3 and 4, section 2). Farther northwestward, reflecting horizon III is continuously traceable over the entire arch of the Shatsky Swell and its northeastern flank toward inner areas of the Tuapse trough. At the steep bent of the southwestern flank, the top of the Cretaceous sequence grades into the erosion surface that is designated in time sections by a letter U (unconformity).

This unconformity is best manifested in seismic profiles crossing the steep 8- to 10-km-high slope of the West Black Sea Basin adjacent to the Crimea (Fig. 3). In the upper part corresponding to the continental slope

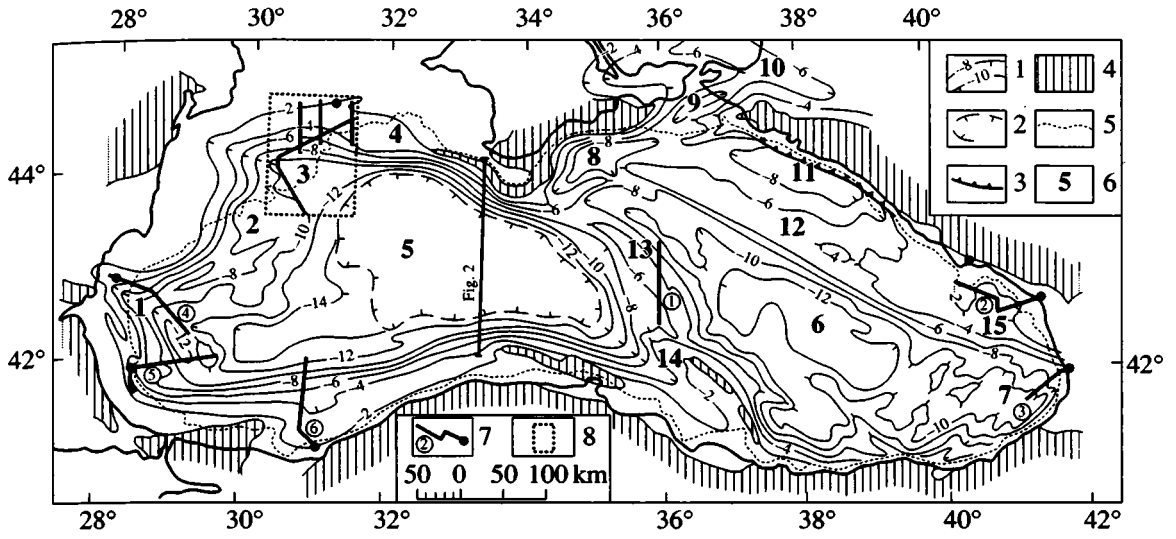

Fig. 3. Structural map of the Cenozoic sediments base in the Black Sea basin (reflecting horizons III and U): (1) isolines (km) at the surface of Cretaceous or eroded Mesozoic deposits; (2) areas of undetected reflections from the Mesozoic surface; (3) Near–Caucasus upthrow–thrust; (4) areas lacking Cenozoic sediments; (5) shelf break; (6) second-order structures: (1) Nizhnekamchiiskii trough, (2) Bulgarian scarp, (3) Istriiskii trough, (4) marginal scarp, (5) West Black Sea Basin; (6) East Black Sea Basin, (7) Guriya trough, (8) Sorokin trough, (9) Kerch-Taman trough, (10) Indol-Kuban trough, (11) Tuapse trough, (12) Shatsky Swell,(13) Andrusov Swell, (14) Arkhangel'skii Swell, (15) Gudauta and Ochamchire arches); (7) lines of seismic profiles whose schematic time sections are shown in Figs. 4 and 5; (8) area of correlated seismic time sections shown in Fig. 6.

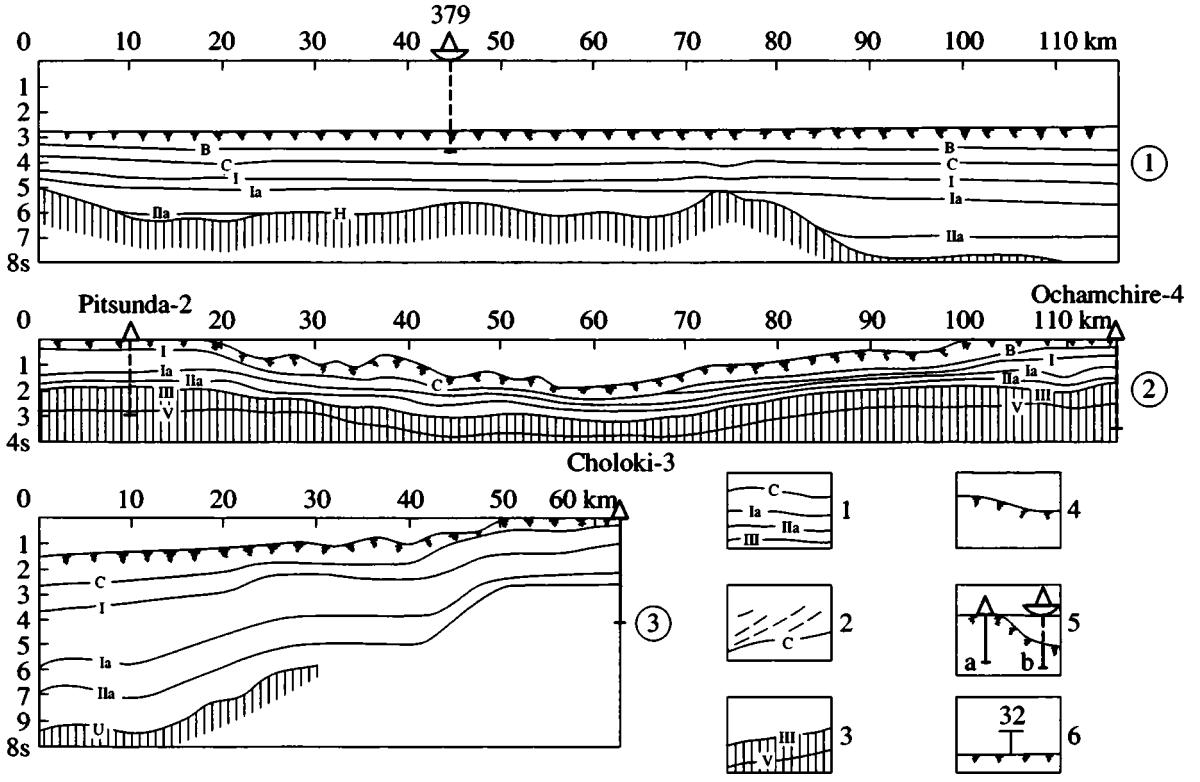

Fig. 4. Schematic time sections of CMP seismic profiles across the East Black Sea Basin and their correlation with borehole sections (profile lines and borehole sites for this and Figs. 5–6 are shown in Figs. 1, 3, 7, 10); (1) key reflecting horizons and their indices: (B) within the Quaternary sediments, (C) the base of the Quaternary, (I) the base of the upper Miocene, (Ia) the surface of the lower Miocene (top of the Maikop Group), (IIa) Eocene surface, (III) the surface of Cretaceous deposits (at slopes and basin bottom grades into reflecting horizon U–eroded surface of Mesozoic deposits), (V) the surface of Jurassic deposits; (2) reflecting horizons of delta clinoforms in the Quaternary sequence; (3) Mesozoic deposits; (4) seafloor; (5) onshore or shelf (a) and deep-sea (b) boreholes; (6) intersections of seismic profiles.

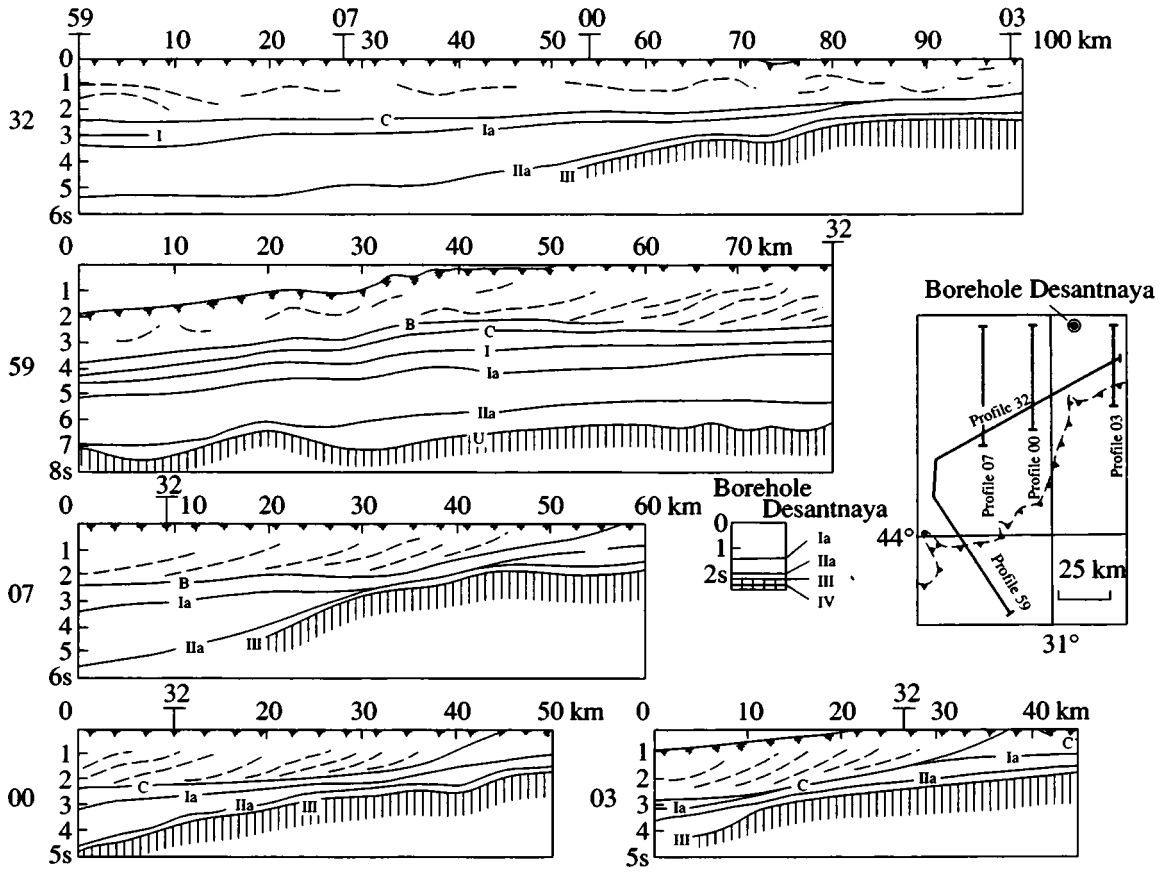

Fig. 5. Schematic time sections of CMP seismic profiles across the Danube River delta and fan in the northwestern shelf of the Black Sea (symbols as in Fig 4).

of the present-day deep-sea basin, there are exposed Cretaceous, Jurassic, and Upper Jurassic–Lower Triassic deposits (the Mesozoic Tavriya Group). West of the Mountainous Crimea, the slope of the basin gradually becomes less steep. In this area, the reflecting horizon III traced in the northwestern shelf and correlated through sections of borehole sections with the upper Cretaceous boundary gradually inclines southward turning into the erosion surface U (Fig. 5).

In the Bulgarian shelf, the reflecting horizon III distinguishable in a thick grid of CMP profiles is reliably correlated with the upper boundary of the Cretaceous rocks in onshore borehole sections (Dachev *et al.*, 1989). Being slightly below the hole toe at the Samotino-more site, the horizon submerges at low angle along the axis of the Nizhnekamchiiskii trough to the West Black Sea Basin, where it grades into reflecting horizon U (Fig. 6, section 4). Southward, the top of the Cretaceous section is penetrated in the Turkey shelf by the Borehole Karadeniz. Extrapolated from the Borehole Karadeniz to Borehole Igneada, the horizon in question is further traceable along the CMP profile into deep sea area, where it joins the reflecting horizon U (Fig. 6, section 5). In the Anatolian shelf of Turkey, the Borehole Akchakhodzha-1 also penetrated the upper Cretaceous boundary correlative with the reflecting horizon III recorded in the CMP profile 10 km north of the borehole near the shelf edge (Fig. 6, section 6). Toward the central part of the basin, it rapidly turns into the reflecting horizon U.

The reflecting horizon III is established only in the mentioned areas of the Black Sea. In the deepest sea areas, the Mesozoic basement surface cannot be identified with the upper boundary of the Cretaceous sequence. At steep slopes, that surface is clearly eroded. This is also evident from structural patterns of overlying Cenozoic formations, which decrease in thickness and pinch out upward the slope (Tugolesov *et al.*, 1990; Gorshkov *et al.*, 1993). The erosional type of boundary is evident in the arch of the Andrusov Swell as well (Fig. 4, section 1), where, in addition to the onlap of overlying Paleogene sediments, several reflecting horizons within the Mesozoic basement of the swell are unconformably truncated by the Mesozoic surface. At present, stratigraphic correlation of these reflecting horizons is ambiguous. At the base of sediments in the West and East Black Sea basins, the reflecting horizon U is noticeably less distinct in the seismic records than at the basin slopes. In the uneven, slightly undulating U surface, there are distinguishable several uplifts of dif-

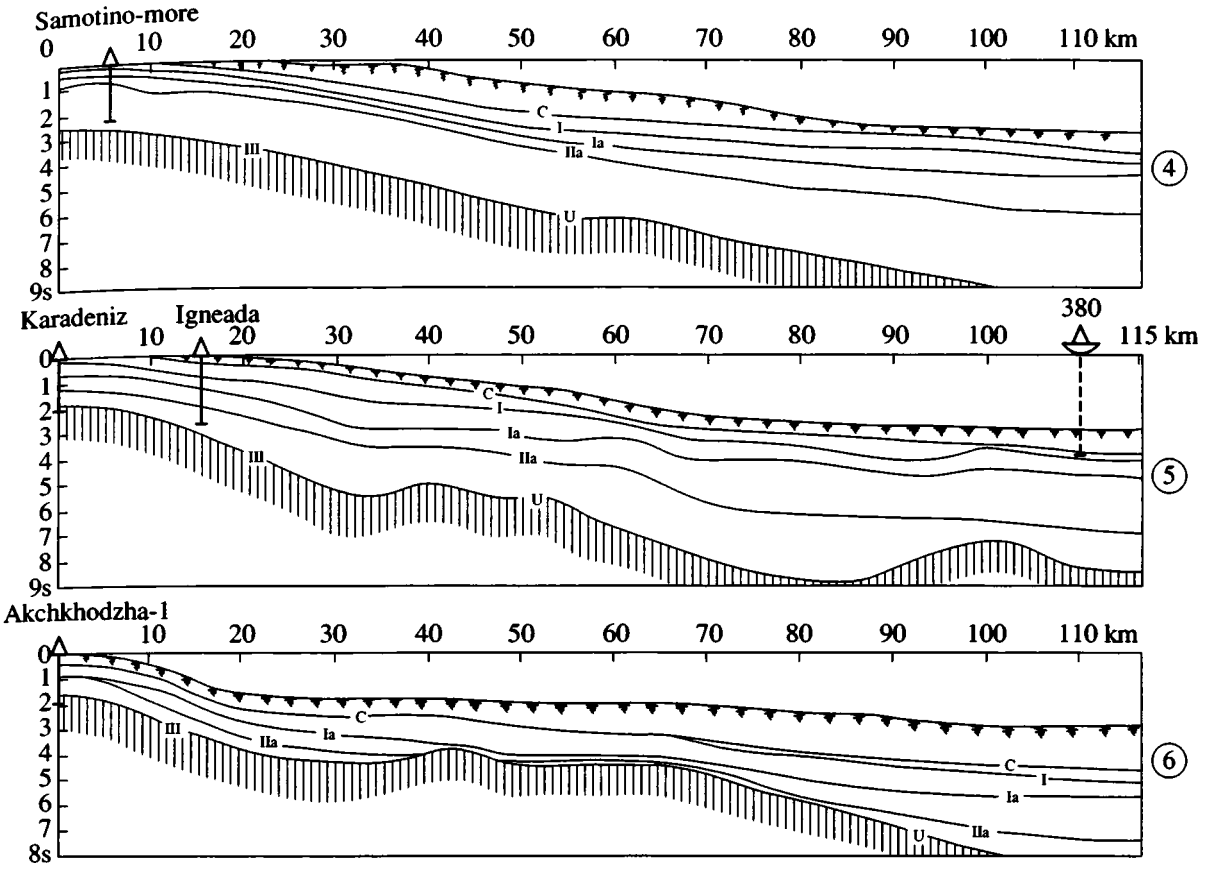

Fig. 6. Schematic time sections of CMP seismic profiles across the West Black Sea Basin and their correlation with borehole sections (symbols as in Fig 4).

ferent size and amplitude. The surface is frequently traceable only owing to the onlap of separate reflecting horizons of the overlying Paleocene–Eocene sequence. The onlap of Paleocene–Eocene sediments whose strata frequently decrease in thickness above uplifts of the reflecting horizon U imply that large undulations of the upper Mesozoic surface, which represent basement of basins, formed in the course of the basement downwarping during the Paleogene. The West Black Sea Basin subsided more intensely than the East Black Sea Basin, and seismic records do not depict the surface U in the deepest part of the former (Fig. 3).

Deeper reflecting horizons are recorded and stratigraphically correlated in the northwestern and Bulgarian shelves. The reflecting horizon IV corresponding to the Jurassic surface is established only in abyssal area of the Georgian segment of the sea (Fig. 4, section 2). Northwestward, it is traceable along the Shatsky Swell axis up to 38°E, but farther it is impossible to identify in available records the reflecting horizons within Mesozoic sequences.

REFERENCE REFLECTING HORIZON IIa (TOP OF THE EOCENE)

The reflecting horizon corresponding to the top of the Eocene sequence was first established in the Georgian segment of the sea (Fig. 7). At the beginning, it was not discriminated from the reflecting horizon corresponding to the Cretaceous surface because of a small thickness and uniform carbonate composition of Paleocene–Eocene and Cretaceous rocks (Goncharov *et al.*, 1972). Later on, key reflecting horizons IIa and III distinguishable in this area were continuously traced throughout the entire northeastern flank of the Shatsky Swell and correlated with boundaries penetrated onshore by boreholes Ochamchire-4 and Pitsunda-2 (Fig. 4, section 2). In this large area, the reflecting horizon IIa is very distinct in all seismic profiles above the acoustically opaque sequence. Reflecting horizons IIa and III are persistently conformable in the records and closely spaced because of a small thickness of Paleocene–Eocene sediments.

In the northwestern and Bulgarian shelves, reflecting horizon IIa correlative with the top of Eocene strata in onshore and shelf boreholes is also easily identifiable

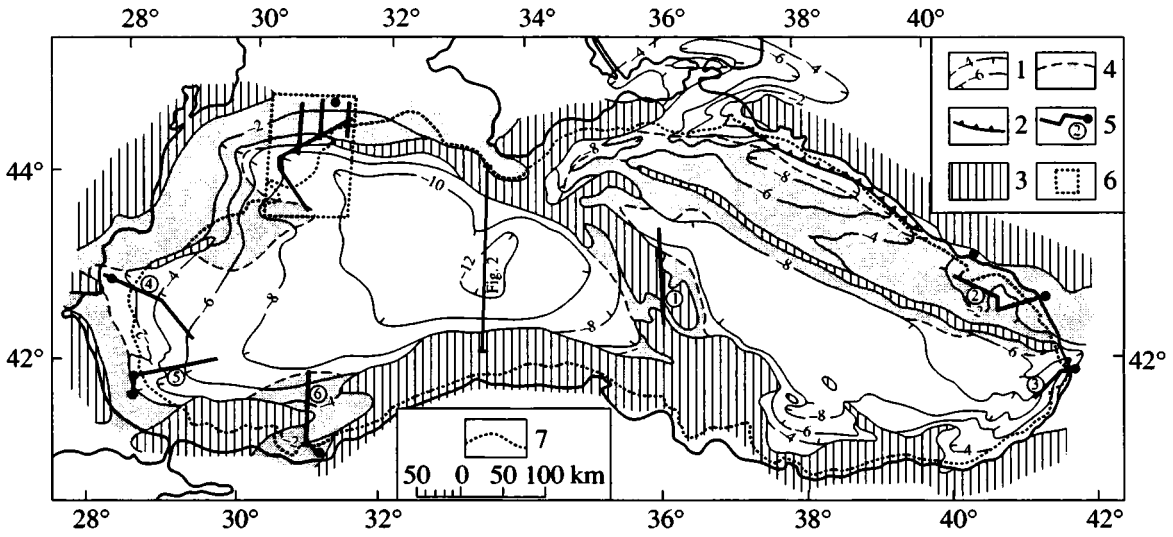

Fig. 7. Map of reflecting horizon IIa (surface of Eocene sediments) in the Black Sea basin: (1) isobaths at the surface of Eocene sediments, km; (2) zone of faults crossing the surface of Eocene sediments; (3) areas lacking Paleocene–Eocene sediments; (4) areas with low thickness of Paleocene–Eocene sediments (closely spaced reflecting horizons IIa and III); (5) lines of seismic profiles whose schematic time sections are shown in Figs. 4 and 6; (6) area, where time sections of seismic profiles shown in Fig. 5 are correlated; (7) shelf break.

in seismic records. The Paleocene–Eocene carbonate facies are also of a low thickness here (Fig. 5).

In distinction from other areas of the northern Bulgarian shelf, the thickness of Paleocene–Eocene sediments significantly increases in the Nizhnekamchiiskii trough. Reflecting horizon IIa correlated with the Eocene surface in the Borehole Samotino-more is continuously traceable in seismic profiles extending in abyssal basin. All along, the reflecting horizon is easily distinguishable even among multiple reflections (Fig. 6, section 4).

Southward, reflecting horizon IIa is also well traceable from the shelf Borehole Igneada to abyssal basin (Fig. 6, section 4). It is more arbitrarily traced in the Anatolian shelf of Turkey away from the Borehole Akchakhodzha-1, because thickness of Paleocene–Eocene sediments over the greater slope part of the West Black Sea Basin is low. Sometimes, these sediments pinch out and the Eocene surface is recorded vaguely (Fig. 7). Farther toward the basin, reflecting horizon IIa becomes again distinct.

In the central part of the West Black Sea Basin, south and southwest of the Crimea, reflecting horizon IIa is located particularly deep, at depths of 10–12 km below seafloor. Because of the horizontal attitude of sedimentary beds in this area, it is difficult to trace this horizon among multiple reflections. It is distinguishable owing to rootless dome-like uplifts of the Maikop sequence with the elevated gas concentrations. The high gas content in sediments decelerates the velocity of seismic waves within these uplifts, and the reflecting horizon IIa is seen therefore in the seismic record in a form of contraflexures, the so-called "velocity synclines," which are distinct in multiple reflections below these structures (Gorshkov *et al.*, 1993; Meisner *et al.*, 1996). Owing to this, the reflecting horizon IIa is rather confidently correlated between these seismic record features.

The thickness of Paleocene–Eocene sediments in the West Black Sea Basin is particularly high, exceeding 5 km in most subsided areas. Two-three additional reflecting horizons, similar in acoustic properties to the reflecting horizon IIa, but less persistent, are distinguished within this sequence. At slopes of the basin, the thickness of Paleocene–Eocene sediments gradually decreases, reflecting horizons rise upward, become closely spaced, and pinch out overlapping the Mesozoic surface. The reference reflecting horizon IIa similarly rises up and overlaps the Mesozoic surface.

In a similar manner, the Paleocene–Eocene deposits occur at slopes of the Andrusov Swell, where they completely pinch out near the swell axis (Figs. 7 and 4, section 1). Nevertheless, the reflecting horizon IIa is so peculiar here that it is confidently recognized as identical on both sides of the swell, in the West and East Black Sea basins. In addition, it is almost continuously traced in both basins along a shallow depression between Andrusov and Arkhangelskii swells (Fig. 7).

In the East Black Sea Basin, the thickness of Paleocene–Eocene sediments is substantially lower, slightly exceeding 3 km in its axial part. The surface topography of the Mesozoic basement is relatively intricate here and, correspondingly, the thickness of the Paleocene–Eocene sequence is laterally very variable. The reflecting horizon IIa is traceable here with a higher confidence than in the West Black Sea Basin.

Along the slope between the East Black Sea Basin and Shatsky Swell, the sediments underlying the reflecting horizon IIa pinch out at depths of 1.5–2 km below the upper slope edge. Above the edge, the reflecting horizon similar in seismic characteristics is again developed at the swell arch. Locally, it is bent and traceable for some distance down the slope. The study of the slope structures showed that slopes of two Black Sea basins developed as huge synsedimentary flexures, in subsided limbs of which (in basins) the thickness of sediments is several times greater than in the risen limbs (i.e., in adjacent shelves, swells, and arches). By the limbs junction (i.e., in basin slopes proper), the thickness of every formation decreases up to their complete disappearance (Tugolesov *et al.*, 1985, 1990; Gorshkov *et al.*, 1993).

In addition to these structural correlations, the reflecting horizon IIa is approximately correlated in the Guriya trough with the section of the onshore Borehole Choloki-3 (Fig. 4, section 3). The Guriya trough joins in fact the East Black Sea Basin and the reference reflecting horizon IIa is continuously traceable within this area.

REFERENCE REFLECTING HORIZON Ia (THE LOWER MIOCENE SURFACE)

The first lithological unit with defined stratigraphic position in the Black Sea was the Maikop Group (Oligocene–lower Miocene). By analogy with the on-land Kerch–Taman trough, folded zones discovered beneath the sea floor were regarded as characteristic of thick clayey Maikop sediments in the Tuapse and Sorokin troughs. In the West Black Sea Basin, similar indicative feature were numerous dome-like uplifts and related single mud volcanoes discovered south of the Crimea (Garkalenko *et al.*, 1976). The detailed survey of 1980s discovered additional mud volcanoes and, later on, they were found in the Sorokin trough. The study of mud volcano breccia showed that it is mainly composed of the Maikop clay (*Mud volcanism...*, 1994; Meisner *et al.*, 1996). Thus, the original inferences based on structural data were confirmed.

The surface of the Maikop deposits (reflecting horizon Ia) was arbitrarily detected in seismic profiles across the Taman shelf, which were approximately correlated with on-land sections, and the surface was traced farther to the deep areas. Later on, the Borehole Rifovaya-302 (Figs. 1, 8) was drilled in the Taman shelf at the arch of the eponymous anticline. The surface of the Maikop Group penetrated by this borehole at depth of 600 m coincides with the defined reference reflecting horizon Ia with uncertainty of several tens of meters (several milliseconds in time record).

Reflecting horizon Ia is horizontal almost throughout the entire deep area of the sea, being inclined under very low angle toward central parts of the West and East Black Sea basins. In deepest parts of both basins, it is either shaded in seismic records by multiple reflections being located slightly deeper than "second floor" (the multiple record of sea bottom) or almost coincides with the latter (Fig. 8). In the elevated slope areas of the basins and in the arch and limbs of the Andrusov Swell, where the reflecting horizon Ia is above the "second floor," it is underlain by acoustically transparent facies. These peculiar seismic records are characteristic of the Maikop deposits (it should be noted that they have not been observed in anticlinal folds of the Tuapse and Sorokin troughs, where seismic patterns in the Maikop sequence are similar to those in underlying strata).

From the areas of acoustically "transparent" Maikop sequence, reflecting horizon Ia is extrapolated to central areas of the basins, where it is approximately drawn among multiple reflections. In the East Black Sea Basin, is it traceable everywhere with a sufficient confidence almost immediately below the multiple reflections of the "second floor." Its position is confirmed by correlation with the section of the onshore Borehole Choloki-3 (Fig. 4, section 3). In the West Black Sea Basin, it is more difficult to discriminate the reflecting horizon Ia among multiple reflections except for the seismic record obtained in the western part of the basin (Figs. 6, 8) and northwestern shelf (Fig. 5), where this reflecting horizon is above the "second floor."

In the Ochamchire and Gudauta arches and in the arch of the Shatsky Swell, where sediments of the Maikop Group are of a low thickness, reflecting horizon Ia that is almost horizontal and conformable to reflecting horizon IIa is correlated with sections of the onshore Boreholes Ochamchire-4 and Pitsunda-2 (Fig. 4, section 2). Northeastward, it plunges at a low angle toward the Tuapse trough. The thickness of the Maikop sequence increases in this direction, although the reflecting horizon Ia remains continuous, retaining its characteristics up to first frontal anticlines in the trough.

In folded zones, key reflecting horizons are traceable only along axes of those synclines, which are open toward the areas with horizontal bedding of Oligocene–Miocene sediments. Based on several synclines, one can outline positions of reflecting horizons Ia, I, and C in sections of anticlines, and then it is possible to extrapolate them over the entire fold zone of the trough. Such a procedure is performed now only for reflecting horizon I (Meisner and Tugolesov, 1998). Extrapolation is more difficult for the lower reflecting horizon Ia, because the numerous high-frequency seismic profiles across the Tuapse trough penetrated shallower into sediments, horizon Ia is outside the records in many areas. The Sorokin trough has not been studied in detail, and all key reflecting horizons are shown in its structural maps very schematically.

It should be added that the Maikop sediments bounded at the top by reflecting horizon Ia pinch out over a significant part of the southwestern slope of the Shatsky Swell and appear again in the arch of the struc-

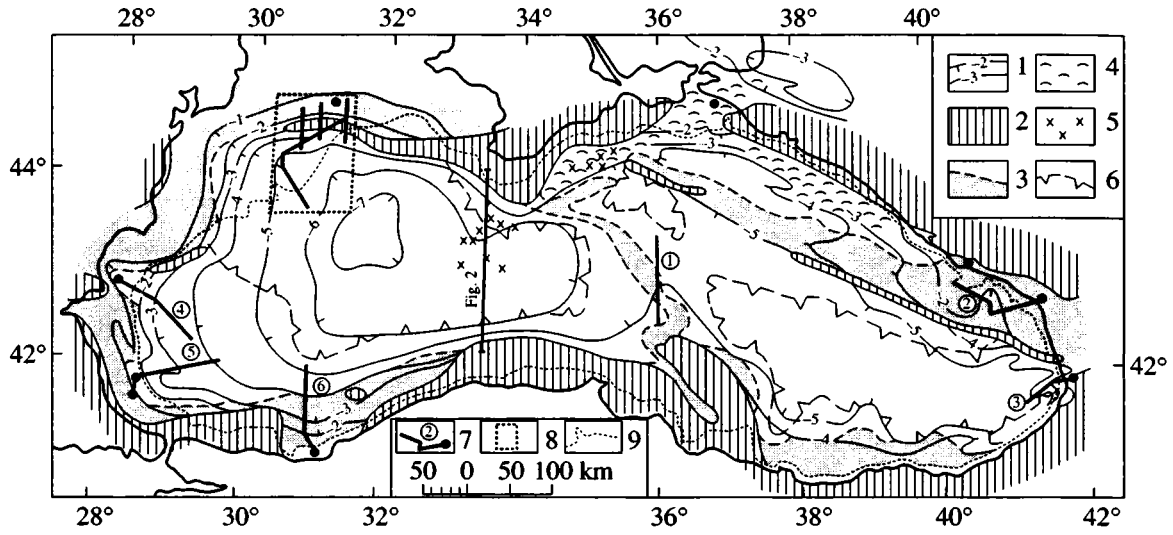

Fig. 8. Map of reflecting horizon Ia (surface of the lower Miocene) in the Black Sea basin: (1) isobaths at the lower Miocene (Maikop Group) surface, km; (2) areas lacking lower Miocene (Maikop Group) sediments; (3) areas with low thickness of lower Miocene (Maikop Group) sediments(closely spaced reflecting horizons Ia and IIa); (4) fold zones within Maikop clays in troughs; (5) underwater mud volcanoes whose breccia consists mainly of Maikop clays; (6) areas, where the reflecting horizon Ia is masked by multiple reflections; (7) lines of seismic profiles whose schematic time sections are shown in Figs. 4 and 5; (8) areas, where time sections of seismic profiles shown in Fig. 5 are correlated; (9) shelf break.

ture, where they have a reduced thickness (Fig. 8). Thus, the synsedimentary development of flexure bridging the Shatsky Swell and East Black Sea Basin commenced, in our opinion, in the Paleocene–Eocene and was also in progress during Oligocene–early Miocene. The junction limb of the flexure was subjected to erosion at that time. Similar erosional processes are characteristic of the present-day continental slopes.

REFERENCE REFLECTING HORIZON I (THE UPPER MIOCENE BASE)

Sediments overlying the Maikop Group produce abundant, virtually horizontal reflectors in seismic records. The most distinct and continuous among them is reflecting horizon C correlated with the base of the Quaternary sequence. Another strong reflecting horizon (I) is distinguished between the latter and reflecting horizon Ia (Fig. 9).

In the time section of Profile 18163 (Fig. 5, section 5), reflecting horizon I was correlated with the base of the Pliocene in the shelf Borehole Igneada. It should be noted that Foos and Manheim (1975) described stratigraphic subdivision of the borehole section without any substantiation. The reflecting horizon I proper is traced in the time section arbitrarily. Its correlation with the section of DSDP Hole 380, where it is drawn at a depth of approximately 1000 m below sea floor, is also ambiguous. In the western part of the sea, where the thickness of sediments sandwiched between reflecting horizons I and C is several times lesser (*Al'bom strukturnykh...*, 1989), identification of reflecting horizon I in time sections is approximate. It is probable that sedimentary sequence is lacking here several Pliocene and upper Miocene formations and reflecting horizon I turns out to be at a lower stratigraphic level. Therefore, in all time sections of seismic profiles available for this area, reflecting horizon I is either drawn arbitrarily, to a significant extent, or not defined at all (Figs. 5, 6).

The reflecting horizon under consideration is actually extrapolated to the Kerch-Taman shelf from the position of the Sarmatian limestones upper surface in the adjacent land. Being extrapolated then to deep-sea areas beyond the folded zone, this reflecting horizon is traced along the arch of the Shatsky Swell, where its intensity in seismic records is slightly different than that of neighboring reflecting horizons. In the fold zone of the Tuapse trough, reflecting horizon I is defined in high-frequency CMP profiles along axes of synclines. Farther southeastward, in the Gudauta and Ochamchire arches, this reflecting horizon is correlated, at first approximation, with the base of the Maeotian regional stage penetrated onshore by boreholes Pitsunda-2 and Ochamchire-4 (Fig. 4, section 2).

In southwestern Georgia adjacent to the Black Sea, the Maeotian sediments overly rocks of different ages, the Cretaceous strata included. The thickness of Maeotian–Pliocene sediments in the southern Kolkhida depression is much lower than in the Ochamchire arch area. Borehole Lesa-1 penetrated at depth of 2.5 km the base of the Maeotian sequence. Borehole Maltakva-3 3.5 km deep did not reach this level, and borehole Supsa-32 reached it at depth of 4 km. Nevertheless, reflecting horizon I arbitrarily drawn in the Guriya trough appears to be close to the base of a thick Maeo-

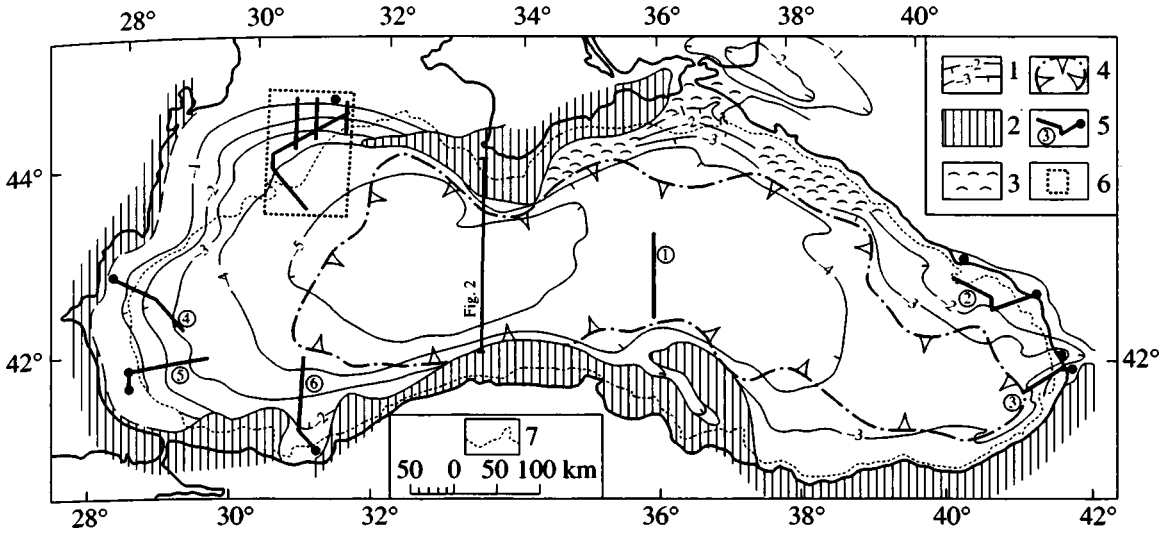

Fig. 9. Map of reflecting horizon I (base of the upper Miocene) in the Black Sea basin: (1) isobaths of the upper Miocene base, km; (2) areas lacking upper Miocene–Pliocene sediments; (3) fold zones within Neogene sediments in troughs; (4) contour of area with continuous correlation of the Reflecting horizon I in the deep-sea basins; (5) lines of seismic profiles whose schematic time sections are shown in Figs. 4 and 6; (6) area, where time sections of seismic profiles shown in Fig. 5 are correlated; (7) shelf break.

tian–Pliocene sequence penetrated by boreholes. In the seismic profile stretching along the southeastern side of the Guriya trough, it is also close to the base of the Maeotian sequence penetrated onshore by borehole Choloki-3 (Fig. 4, section 3).

Thus, we correlated fragments of reflecting horizon I beginning from the Georgian coast westward, i.e., in direction opposite to the successive series of profiling tracks performed in the Black Sea basin from the west to the east. It was established that the reflecting horizon I is traceable confidently and continuously in all available seismic profiles throughout the most deep areas of the sea (Fig. 9). In peripheral parts of the abyssal basins, discrimination of the reflecting horizon is difficult because it is discrete and poorly manifested sometimes. Nevertheless, it is approximately correlative with the Maeotian–Sarmatian boundary in all shelf and onshore boreholes mentioned above. In other words, it is located near the base of the upper Miocene sequence.

REFERENCE REFLECTING HORIZON C (THE QUATERNARY BASE)

The reflecting horizon C defined in first CMP seismic profiles south of the Crimea at the beginning of the 1960s (Goncharov *et al.*, 1972) was traced by 1975 almost up to DSDP Hole 380/380A drilled in the same year.

According to initial stratigraphic subdivision of the section, the base of the Quaternary was placed at depth of about 840 m below sea floor (Ross *et al.*, 1975), which corresponded approximately to the occurrence depth of reflecting horizon B. The initial stratigraphic scheme was revised after subsequent core study. Based on abundant, though endemic diatoms, A.P. Jouse and V.V. Mukhina determined the base of the Quaternary at a depth of 660 m and referred the underlying sequence, down to 840 m, to the Pliocene and Pontian (*Geologicheskaya istoriya...*, 1980; *Initial Reports...*, 1978). In 1978, the regular seismic survey that gradually advanced from western to eastern areas was initiated. E.M. Khakhalev and his colleagues who carried out the survey accepted the new stratigraphic subdivision of section penetrated by Hole 380/380A and established that reflecting horizon B poorly traceable in seismic sections near the hole approximately corresponds to the newly determined base of the Quaternary.

When all deep-sea areas were covered by a dense system of CMP seismic profiles, the reference reflecting horizon C was correlated confidently and continuously almost throughout the Black Sea (Fig. 10). The reflecting horizon B is also traceable continuously, but in noticeably smaller areas. For instance, it is obscure in the vicinity of DSDP Site 380.

In order to exactly calibrate seismic time sections, DSDP Site 380/380A was crossed by several CMP (Fig. 1) and high-resolution seismic profiles, the latter with 1 s travel time, not deeper. The diagram of seismic wave velocity variations with depth was plotted for the site area with due account for all available data on lithology, acoustic velocity in cores samples, etc. Based on the diagram, the hole section was recalculated for the time scale and graphically juxtaposed with the seismic time section (Gorshkov *et al.*, 1993).

The Hole 380/380A section has a very distinct lithological boundary. At a depth of 840 m below sea floor, this boundary separates the sequence of different oozes, soft clays, marls, and lacustrine chalk with abundant

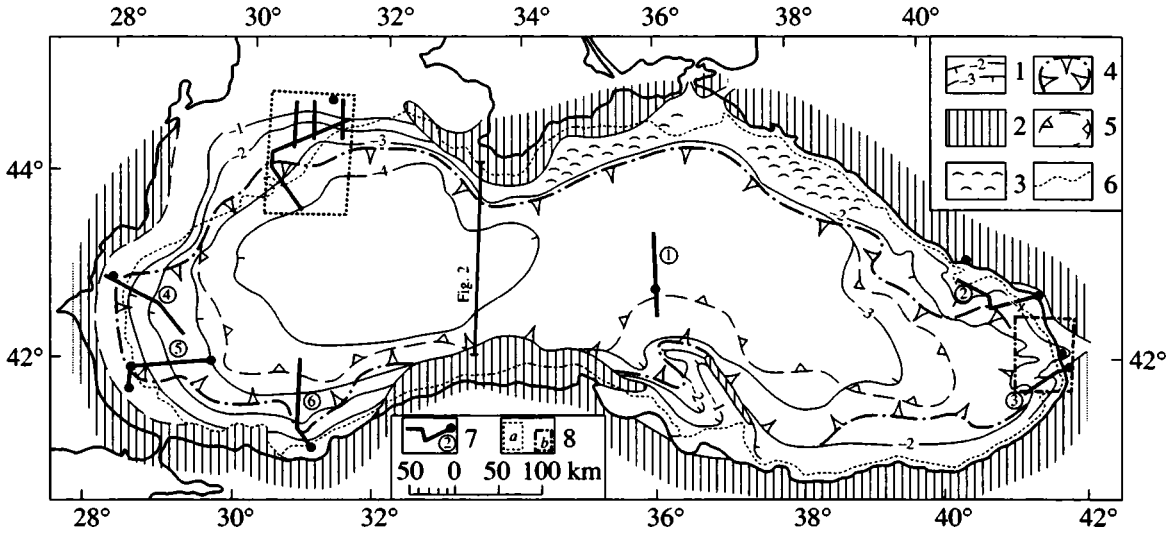

Fig. 10. Map of reflecting horizons C (base of the Quaternary) and B (within the Quaternary sediments) in the Black Sea basin: (1) isobaths of the of the Quaternary base, km; (2) areas lacking or of thin Quaternary sediments; (3) fold zones within Quaternary sediments in troughs; (4) contour of area with continuous correlation of the reflecting horizon C in the deep-sea basins; (5) the same for the reflecting horizon *B*; (6) shelf break; (7) lines of seismic profiles whose schematic time sections are shown in Figs. 4 and 6; (8*a*) area, where time sections of seismic profiles shown in Fig. 5 are correlated; (8*b*) segment of the eastern Black Sea shown in Fig.11.

diatoms from the laminated black clays gradually compacting downward and lithified. This boundary almost coincides with the strong reflecting horizon C well traceable in all profiles. In the interval of 864 to 883 m, clays are underlain by landslide breccia composed of marl and limestone fragments with stromatolites, which are cemented by calcareous, mainly dolomitic material. Downward to the hole base, there are similar dark lithified clays alternating with thin-bedded marls, limestones, and dolomites.

When the reference reflecting horizon C was mapped throughout the entire Black Sea basin, the revealed additional features showed its probable confinement to the base of the Quaternary. In the northwest, for instance, the reflecting horizon C underlies a thick sequence of clinoforms corresponding to underwater fans of the Danube River avandelta (Fig. 6). The formation of this sequence commenced most likely in the Quaternary only, because the Danube River discharged sedimentary material in the spacious Outer Carpathian lake until the end of the Pliocene (Oncesku, 1960). Pliocene sediments in the Carpathian foredeep are very thick, whereas the Quaternary sequence is of insignificant thickness here. It is likely, that the Danube River sediment discharge found the way to the Black Sea across the Carpathian foredeep completely compensated by sediments only at the beginning of the Quaternary.

Similar position is characteristic of the reflecting horizon C in the Don-Kuban paleodelta area, where it also at the base of thick clinoforms of a large fan, the second one in size after the Danube River fan (Tugolesov, 1994). Characteristic of both areas is also a similar development history: the Indol-Kuban trough with a thick upper Miocene–Pliocene sequence (Gorshkov *et al.*, 1989) was also compensated by sediments by the beginning of the Quaternary, when sediment discharge of the Paleo-Kuban and, partly, Paleo-Don rivers reached the Black Sea.

In western Georgia, the southern part of the Kolkhida depression that intensely subsided during the late Miocene–Pliocene continued to subside also in the Quaternary. The Borehole Maltakva-3 touched the base of the Quaternary at a depth of 1300 m, whereas Borehole 89 in the Supsa anticline area did not reach it even at the depth of 1500 m (the hole toe). Offshore, the reflecting horizon C is arbitrarily traceable near the Georgian coast (Fig. 10). Nevertheless, it was mapped along the system of profiles correlated between each other rather confidently. It appears that the horizon depth offshore closely corresponds to the base of the Quaternary recovered by onshore boreholes (Fig. 11).

Continuous tracing of the reflecting horizon B from DSDP Hole 379 almost to Hole 380 (Fig. 10) also indicates that the base of the Quaternary in the Hole 380/380A section is placed too high by Jouse and Mukhina. Drilling in Hole 379 stopped in the Quaternary sediments, but the reflecting horizon B is observed in its vicinity near the hole toe (Fig. 4, section 1).

It should be stressed that stratigraphic subdivision of the Hole 380/380A section was insufficiently substantiated. For instance, D. Ross notes that available paleontological materials are insufficient for a reliable subdivision of the drilled section (*Initial Reports...*, 1978, p. 17). H.-J. Schrader (ibid., p. 850) and V.N. Semenenko (1984) believe that endemic diatom species

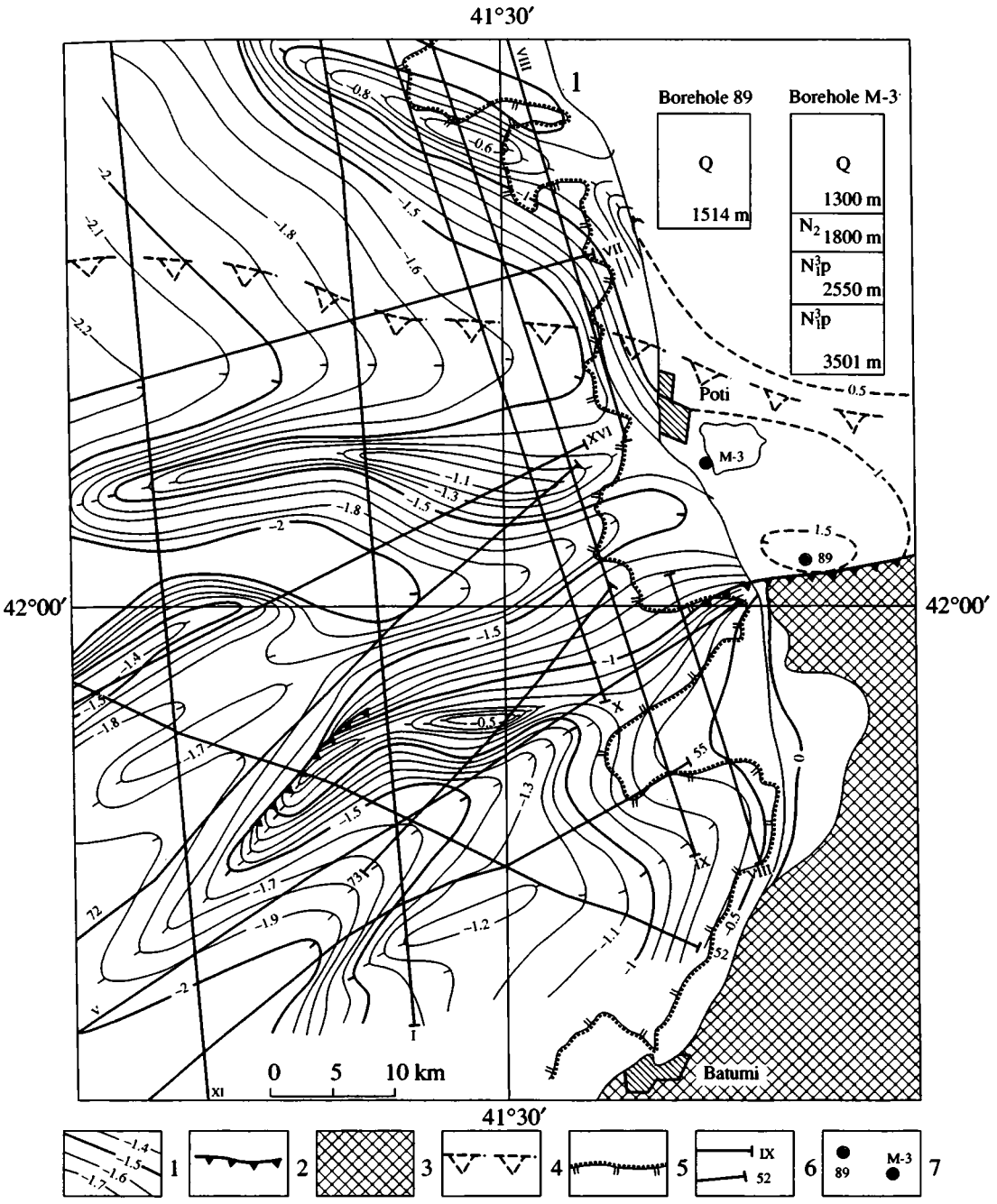

Fig. 11. Structural map of the reflecting horizon C (base of the Quaternary) in the eastern Black Sea (for area location see Figs. 1 and 10): (1) isobaths of the reflecting surface C, km; (2) thrusts (denticles oriented toward the overthrust wall); (3) exposures of pre-Quaternary sediments; (4) edge of the Paleocene–Miocene slope of the basin; (5) present-day shelf-break; (6) CMP seismic profiles and their numbers; (7) boreholes.

found in cores give no grounds for more or less reliable stratigraphic subdivision of the section. All this leads to the conclusion that the reference reflecting horizon C corresponds to the base of the Quaternary. This conclusion we briefly substantiated long ago (Tugolesov *et al.*, 1985).

Two years later, the results of the study of nannofossils from the core of Hole 380/380A were published (Golovina *et al.*, 1987). Authors of this work believe that nannofossils are the only group suitable for zonal subdivision of the Black Sea deep-water sediments (because planktonic foraminifers are not found and diatoms are represented by endemic species). In the Hole 380/380A section, nannofossils are however more or less abundant only up to depth of 480 m. Below this level, they occur only in the interval of 840–884 m,

being represented by the monotype flora *Braarudosphaera bigelowi*. This species is of a low stratigraphic significance (range from Jurassic to Quaternary), but it is indicative of marine, slightly desalinated environments. Diatoms abundant in the interval from 660 to 860 m show presence of marine forms in the lower part (below 840 m) and rapid replacement by brackish-water and freshwater forms upward the section. Based on the coincidence of data on diatoms and nannofossils, it was concluded that the level of environmental change "acquires significance of a reliable boundary for correlation of deep-sea sediments and on-land sections. It orientates us toward the search of levels in on-land sections, which correspond to similar substantial reorganizations in the Neogene paleogeography of the Eastern Paratethys, and correlation is limited by two variants" (Golovina *et al.*, 1987, p. 43). The first variant of such a level is known on land in the lower part of the Pontian. For instance, the cited authors established the autochthonous marine nannoplankton in Pontian sections of the Taman Peninsula and near Odessa. However, taking into consideration our results, they also outline the second possible variant of correlation. In the Neogene history of the Pontian–Caspian region, there is known yet another level of transition from the marine sedimentation to brackish-water one, namely the Akchagylian–Apsheronian boundary, and the only period of marine diatom algae development, the Akchagylian. It is also known that almost all diatom species found in the interval of 840–884 m are known from the Akchagylian and younger sediments of the Pontian–Caspian region (Golovina *et al.*, 1987).

It is clear that this variant coincides with the stratigraphic position of the reflecting horizon C proposed in this paper.

CONCLUSION

The structural peculiarity of the Black Sea sediments predominantly deposited at the abyssal depth determined specific approach to study and stratigraphic subdivision of thick sedimentary sequence at the bottom. Because of a large depth, information on the geological structure of the basin can be obtained mainly by means of seismic profiling.

On land, stratigraphic subdivision of sedimentary sequences is based on study of rocks from exposures and drill cores. Reference sections of this kind are unavailable for abyssal basins of the Black Sea. On the other hand, the continuity of seismic records is favorable for a detailed structural deciphering of sedimentary sequences and for tracing of individual horizons over large distances.

The suggested stratigraphic subdivision of sedimentary sequences in the Black Sea basin is based on correlation between seismic time sections and between the latter and geological sections recovered by boreholes on land and in various shelf areas. Characteristic structural features revealed in sedimentary cover, such as the underwater fans, diapirs, specific pinch-out zones of sedimentary formations in paleoslopes of deep basins, etc., are used to verify the correlation of key reflecting horizons.

The correlation shows that the thickest and most complete Cenozoic sedimentary cover is characteristic of the abyssal sea areas. Most persistent and continuously traceable key reflecting horizons are detectable in these areas. Accordingly the future strategy of investigations in peripheral areas of the Black Sea should include the seismic units tracing from abyssal parts of the Black Sea to the continental slopes and, when possible, to shelves in order to get a more precise stratigraphic subdivision of Cenozoic sediments.

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Middle Pliocene Small Mammals of the Russian Plain

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Abstract—Materials on small mammals collected during last years enable discrimination of an autonomous Uryv faunal assemblage in the Pliocene succession of mammal communities of the Russian Plain. Data on taphonomy and compositional peculiarities of the assemblage are considered. Evolutionary transformations in communities of the Uryv fauna are established. The fauna evolutionary period corresponds to Zone 16 by P. Mein and spans the greater middle and basal upper Pliocene.

Key words: middle Pliocene, small mammals, Russian Plain.

INTRODUCTION

In their fundamental handbook *Quaternary System*, Vangengeim and Zazhigin (1982) noted the necessity of defining the Uryv faunal assemblage that characterizes the "pre-Khaprov" Pliocene stage in development of mammal communities. The detailed arguments favoring their idea have been also considered.

Recent data most clearly show that discrimination of the Uryv assemblage is an urgent task. The available factual materials point to a significant evolutionary gap between mammal faunas from the Khaprov sediments of the Sea of Azov region and from the older Moldavian assemblage. Substantial morphological differences between the Khaprov and Moldavian mammal assemblages suggested that a long and important paleogeographic stage is missing from the stratigraphic scheme. This is also evident from an intermediate character of the Kvabeb and Kyzyl-Aigir mammalian faunas found in the Caucasus and West Siberia, respectively (Vekua, 1972; Gabuniya, 1986; Zazhigin and Zykin, 1984; Zykin *et al.*, 1987). In addition, the lower Pliocene corresponds in old schemes to a single phase in development of mammal fauna (zone MN 14 by P. Mein), whereas the upper Pliocene spans three zones MN 15, MN 16, and MN 17 (Mein, 1975; Agadzhanyan, 1986). Thus, the former schemes inconsistent with the dynamics of faunal transformation also call into question their correctness.

The detailed study of Piacenzian marine sediments near the town of Gela in Sicily resulted in discovery of an autonomous Gelasian paleogeographic stage that is older than the Calabrian and younger than Piacenzian proper (Berggren *et al.*, 1995; Cita *et al.*, 1996). Consequently, two subdivisions, instead of a single former unit, i.e., the Piacenzian (middle Pliocene) and Gelasian (upper Pliocene), were defined in the second half of the Pliocene. They span the time ranges of 3.6 to 2.6 Ma and of 2.6 to 1.85 Ma, respectively. The Gelasian Stage (N_3^3gl) was approved as the upper stage of the Neogene System (*Postanovleniya MSK...*, 1998). Thus, it becomes necessary to define continental analogues of the Piacenzian and Gelasian. In the Russian platform, the terminal Pliocene age is attributed to the Khaprov alluvial sediments of the Sea of Azov region, which yield mammal remains of the upper Villafranchian. The upper age limit of the Khaprov fauna corresponds to 2.15 Ma (Vangengeim and Pevzner, 2001). On the other hand, sediments in terrace sections of the Don River upper course yield the mammal remains, which are more archaic than the Khaprov fauna but more advanced as compared with the Moldavian mammal assemblage (Krasnenkov and Agadzhayan, 1976). The bone remains occur in alluvial formations dated back to the middle Pliocene and spanning ranges of the upper Matuyama and Gauss chrons in the paleomagnetic scale (*Verkhnii pliotsen...*, 1985; Krasnenkov *et al.*, 1987). The remains can be regarded under name the Uryv fauna. At the Uryv-1 locality, the bone-bearing horizon is recognizable within the very complete and well-studied section. Several kilometers away, there is the Korotoyak locality containing similar faunas. It is possible therefore to characterize the main stage in development of the Uryv fauna, with the related older and younger phases, and to understand relationships of this fauna with the older (Moldavian) and younger (Khaprov s. str.) mammal assemblages. Thus, all the requirements formulated by Gromov (1948) for the definition of faunal assemblages can be satisfied.

MIDDLE PLIOCENE FAUNAS OF SMALL MAMMALS OF THE DON RIVER UPPER COURSE

The right side of the Don River south of Voronezh is of an intricate geological structure. Coastal precipices

are composed there of Devonian limestones grading into Cretaceous rocks downstream (Nikitin, 1957). Near the villages of Uryv, Semiluki, and Korotoyak, these deposits are eroded by the Don River tributaries. The "scours" extend over a distance of several kilometers and are filled with young sediments of the Devitsa, Potudan', and Tikhaya Sosna rivers. The basal Pliocene and early Pleistocene alluvial sequences of the sections are overlain by the Don Till of the early Neopleistocene and by the complex of fossil loess and soil horizons.

Bone Burials near the Uryv Village

One of the most representative Pliocene–Pleistocene sections is exposed near the Uryv Village of the Ostrogozhsk district, Voronezh oblast. Nikitin (1957) was first to describe this section, in which he found the diverse Pliocene "seed flora." In 1963, F.A Kaplyanskaya and V.D. Tarnogradskii found remains of small mammals in the lower alluvial member. Beginning from 1965, the section and geology of the area were studied under guidance of R.V. Krasnenkov (*Verkhnii pliotsen...*, 1985). In 1965–1972, I.M. Gromov collected representative paleontological materials from the lower bone-bearing horizon. Beginning from 1972, I also collected remains of small mammals from this section and bones from the Pliocene lacustrine clays (Uryv-2 locality) and alluvium (Uryv-1, Uryv-3, and Uryv-4 localities), to study the morphology of main groups of small mammals (Agadzhanyan, 1976, 1986).

Cenozoic sediments near the Uryv Village are traceable over a distance of 3 km upstream the Don River up to their contact with the bedrocks. In this area, the thickness of loess and till beds gradually decreases northward from the Potudan' River valley, and lacustrine–alluvial sediments are replaced by red clays of subaerial origin. Most of the bone-bearing horizons of this section are confined to lacustrine–alluvial sediments, which are exposed under till within the area approximately 300 m across, situated upstream of the ferry crossing. The total thickness of loose sediments is 64 m here (Fig. 1). From the top downward, the following beds are recognizable in the section:

(1) Light loam, porous, black, crumbling into nut-shape fragments and representing the humus horizon of the present-day soil (0–0.8 m thick);

(2) Light loam, porous, gray, with carbonate concretions, enclosing three horizons of buried soils in the lower and middle parts (0.8–14.0 m thick);

(3) Sand, medium- to coarse-grained, locally cross-bedded, light yellow (14.0–14.9 m thick);

(4) Clay, soft, massive, cinnamon-red in the upper part and dark gray in the lower part, with fragments of crystalline rocks (14.9–31.6 m thick);

(5) Sand, fine-grained, horizontally bedded, gray-yellow, with gastropod shells (31.6–36.1);

(6) Alternating medium-grained sands and loams, gray, with cross-bedded lenses of gravelly material, which contain remains of small mammals (36.1–37.4 m thick);

(7) Sand, medium-grained, horizontally bedded, with cross-bedded lenses in the upper part, which enclose mud balls, small chalk pebbles, gastropod shells, teeth and bones of small mammals (37.4–44.3 m thick);

(8) Loam, gray, sandy, horizontally bedded, with rare gastropod shells and bones of small mammals (44.3–46.3 m thick);

(9) Sand, gray, fine-grained (46.3–47.7 m thick);

(10) Clay, compact, gray-brown to black, sometimes with horizontal lamination; the rock is locally enriched in compressed and lignified plant remains, the content of which increases downward to form there a horizon of compact peat up to 30 cm thick. Throughout the bed, sediments contain gastropod shells, rodent teeth and bones, as well as gypsum crystals and druses. A thin bed of medium-grained sand is observable at the contact with the underlying stratum (47.7–49.1 m thick);

(11) Loam, compact, gray, highly sandy, showing fine horizontal lamination in the lower part and enclosing abundant carbonate concretions up to 10–15 cm across in the middle and upper parts (49.1–51.9 m thick);

(12) Sand, fine-grained, horizontally bedded, bright yellow to orange in color; in the lower bed part, sand is medium- to coarse-grained, yellow, and cross-bedded. The bedding is emphasized by thin laminae composed of gray clay, mud balls, gravel of Cretaceous rocks, and gastropod shells. The cross-bedded portion encloses phosphorite concretions, redeposited shark teeth, teeth and bones of small mammals. The lower boundary of the bed is uneven, eroded, well sustained along the strike, and emphasized by gravel lentils. (51.9–57.9 m thick);

(13) Sand, medium-grained, horizontally bedded, almost white in the upper part and light green in the lower one; the bed contains shark teeth, and its lower part is below the water level of the Don River (57.9–63.8 m thick).

In this section, Beds 1 and 2 are composed of subaerial upper Pleistocene sediments, Bed 4 corresponds to the Don Till, alluvial sediments of beds 5–9 are of the early Pleistocene age, and Beds 10–12 represent the Pliocene alluvial and lacustrine sediments with signs of soil formation in Bed 11.

The Uryv-1 burial is in Bed 12. Mammal remains are poorly preserved and scarce. Teeth and bones are light-colored (yellow, light brown). Many vole teeth have destroyed dentine, and their enamel loops have funnel-shape cavities in the middle. Teeth roots are usually broken. There are many enamel chips of small incisors. Lower mandibles have not been found, whereas upper mandible bones are scarce. Some bone fragments crumble into a bright ocherous powder. There are rare fish bones, single frog bones, and fragments of tortoise

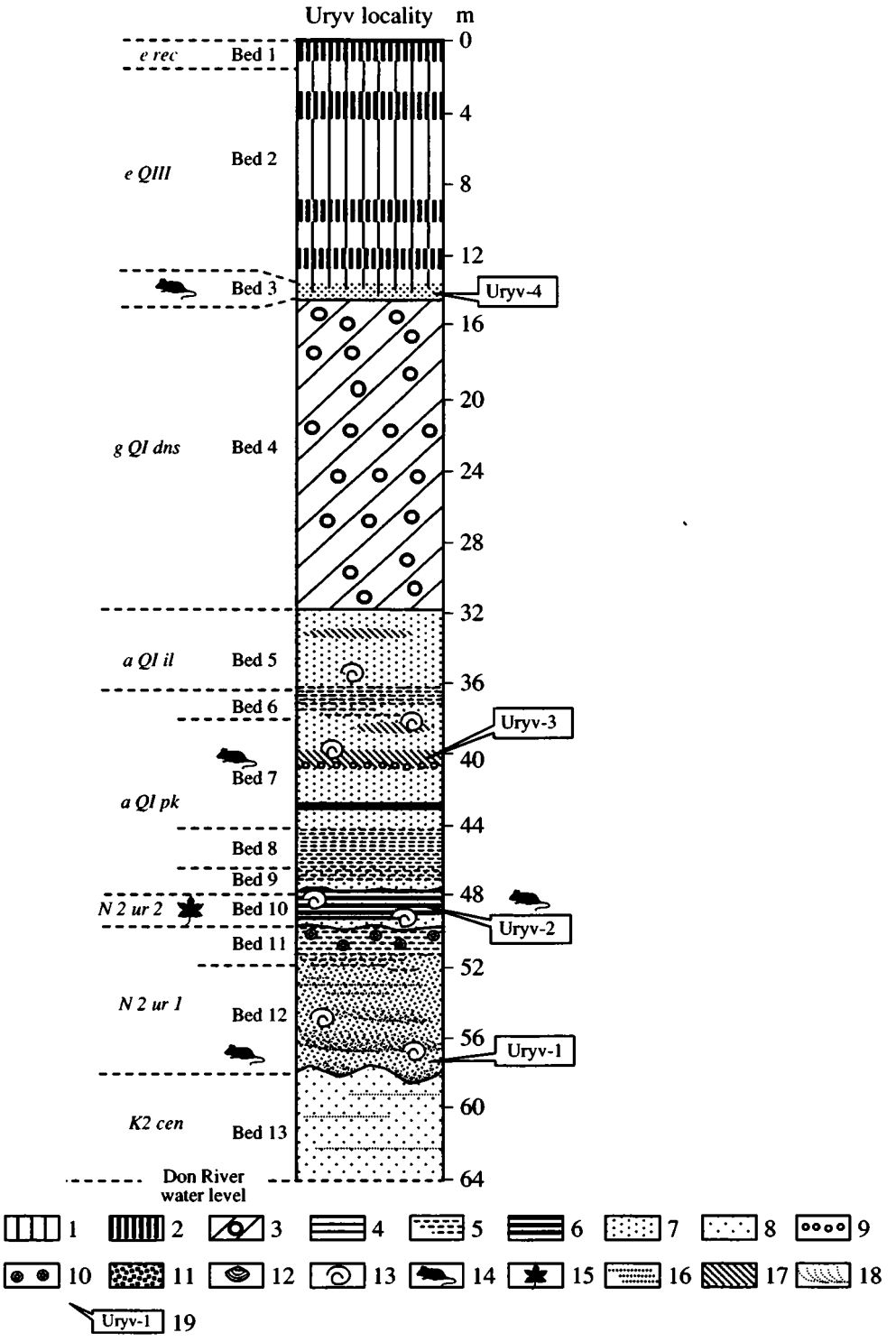

Fig. 1. Lithostratigraphy of the Pliocene–Pleistocene section near the Uryv Village, the Voronezh oblast: (1) loess; (2) humus horizons of recent and fossil soils; (3) clays and loam of the Don Till; (4) horizontally bedded clays; (5) horizontally bedded loam with sand intercalation; (6) compact humus clays; (7) Pliocene–Pleistocene alluvial sands; (8) Cretaceous marine sands; (9) pebble, gravel; (10) carbonate concretions; (11) interbeds saturated with plant organic matter; (12) bivalve shells; (13) gastropod shells; (14) small mammal remains; (15) seeds and plant leafs; (16) horizontally bedded sands; (17) cross-bedded sands; (18) cross-bedding patterns; (19) location of bone-bearing horizons and their names; (D) location of ditches and their numbers; indices of beds after Krasnenkov *et al.* (1987).

mails. In total, over one thousand bone remains are collected from the Yryv-1 locality. The composition of this oryctocoenosis is characterized in Table 1.

Characteristic taxa of the fauna are Lagomorpha and rhizodont voles of the *Promimomys* and *Mimomys* genera. Cericetinae are represented by a peculiar form, which is close to the present-day rat-like hamsters of the Khankai Lowland and Sikhote Alin. Prevalent among voles of the *Promimomys* and *Mimomys* genera are forms with teeth lacking external cement. Only molars of *Mimomys hajnackenais* Fejfar have small cement accumulations in entering corners of teeth. According to the chewing surface structure of M^1 and M^3, development degree of dentine tracks, and number of roots in molars, the vole *Mimomys hajnackenais* Fejfar from the Uryv-1 locality is close to representatives of its type population from the Hajnacka locality (Fejfar, 1961, 1961b, 1964). In addition, it resembles *Mimomys cappettai* from the Balaru-II locality of France (Michaux, 1971).

More than a half of bones collected from the Uryv-1 locality is represented by cement-free teeth of small voles, which have two marks in M^3. The teeth are of two forms: larger *Promimomys (Cseria) gracilis* Kretzoi and smaller *Promimomys (Cseria) baschkirica* Suchov. The former is very similar to the vole described as *Mimomys stehlini* Kormos from Sete and Nimes localities of France and from Escorihuela locality of Spain (Michaux, 1971). The smaller form is most close to *"Mimomys" (Cseria) baschkirica* Suchov from the Villafranchian faunas of Bashkortostan (Sukhov, 1970, 1972). Characteristic of these voles are the well-developed dentine tracks and early disappearance of marks in ontogeny. By its species composition, the assemblage of small mammals from the Uryv-1 locality corresponds to one of the middle Pliocene stages. This is well consistent with the data on geological structure of ancient alluvial sediments in the Don River upper courses (Kholmovoi, 1966; Krasnenkov, 1967, 1968; Krasnenkov *et al.*, 1987). In the evolutionary level of arvicolids, the Uryv-1 fauna is close to that from the middle bone-bearing horizon of the Dniester River terrace IX near the Kotlovina Village of Moldavia (Topachevskii and Nesin, 1989).

The fauna from the Etulia locality in Moldavia looks older, because it is virtually lacking vole teeth with cement (Shevchenko, 1965). Faunas from the Weze locality of Poland (Kowalski, 1960b; Sulimski, 1964) and from Sete, Nimes and Balaru localities of France (Michaux, 1971; Chaline and Michaux, 1966, 1969) are of a comparable age.

Bed 10 of compact dark gray clay represents the Uryv-2 bone-bearing horizon. Remains of mammals from this bed are much better preserved as compared with remains from the Uryv-1 locality. Bones are dark-brown, with a violet tint sometimes. Teeth enamel is colored dark blue to blue. Teeth are almost intact, having unbroken roots and well-preserved growth zones at the base of crowns in young teeth. Common among the bones are fragments of vole lower mandibles and skulls, mandible and teeth of insectivores, bones of birds and fishes, and shells of bird eggs. According to all indications, the mammal remains accumulated in a stagnant, poorly aerated basin undisturbed by currents. The same interpretation of burial conditions of bone remains is inferable from the facies appearance of host alluvial–lacustrine clays with peat lenses. Good preservation of bone fragments and genesis of sediments imply that mammal remains have not been transported to form the taphocoenosis. A thorough examination showed that bones are irregularly distributed in clays of the Uryv-2 horizon. Along the bed strike, they are concentrated in separate "nests," every of which contains bones of one or several individuals. Distribution of material in these "nests" resembles dispersed bird's regurgitates. Birds were probably main agents that transported remains of small mammals. Birds populated probably reed thickets along and in floodplain lakes, and their colonies included gulls, herons, and predatory forms. Amphibians, reptilians, and small mammals were their hunting objects and served as a food for them and their nestlings. Finds of fish, frog, and bird bones in the burial indirectly confirm this inference.

The composition of the Uryv-2 oryctocoenosis is presented in Table 2. The oryctocoenosis appearance is

Table 1. Taxonomic composition of small mammals in the Uryv-1 locality

Insectivora	Abundance	%
Talpa sp.	1	0.08
Megalia sp.	1	0.08
Soriculus sp.	1	0.08
Lagomorpha		
Proochotona ex gr. *eximia-gigas*	7	0.56
Pliolagus brachignatus Kormos	155	12.38
Rodentia		
Myomimus sp.	1	0.08
Cricetulus (Tcherskia) sp.	68	5.43
Pliomys cf. *ucrainicus* Topacevski et Scoric	4	0.32
Villanyia veterior Kretzoi	6	0.48
Promimomys gracilis Kretzoi	35	2.80
Promimomys baschkirica Suchov	32	2.56
Mimomys ex gr. *hajnackensis* Fejfar	48	3.83
Mimomys minor Fejfar	62	4.95
Mimomys sp.	746	59.58
Trogontherium minus Newton	4	0.32
Nannospalax odessanus Topacevski	81	6.47
Total	1252	100.00

Table 2. Taxonomic composition of small mammals in the Uryv-2 locality

Insectivora	Abundance	%
Desmana cf. *senseyi* Kormos	5	0.35
Desmana nehringi Kormos	26	1.84
Talpa ex gr. *minor* Freudenthal	5	0.35
Blarinoides cf. *marinae* Sulimski	19	1.35
Beremendia fissidens (Petenyi)	21	1.49
Sorex ex gr. *runtonensis* Newton	40	2.84
Sorex aff. *drepanosorex* sp.	2	0.14
Sorex sp. A	11	0.78
Sorex sp. B	2	0.14
Sorex cf. *minutus* L.	2	0.14
Sorex sp. indet.	14	0.99
Lagomorpha		
Ochotona sp.	6	0.43
Pliolagus cf. *brachignatus* Kormos	47	3.33
Rodentia		
Dryomys sp.	3	0.21
Apodemus ex gr. *primaevus-jeanteti*	5	0.35
Cricetulus (Tcherskia) cf. *triton* Winston	1	0.07
Baranomys lozyci Kormos	10	0.71
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Villanyia exilis Kretzoi	5	0.35
Promimomys (Cseria) *gracilis* Kretzoi	15	1.06
Promimomys (Cseria) *baschkirica* Suchov	31	2.20
Promimomys sp.	96	6.81
Mimomys ex gr. *polonicus* Kowalski	175	12.41
Mimomys aff. *hintoni* Fejfar	46	3.26
Mimomys pliocaenicus minor Fejfar	11	0.78
Mimomys sp.	801	56.81
Trogontherium sp.	1	0.07
Carnivora		
Mustelidae indet.	1	0.07
Total	1410	100

determined by rhizodont voles *Promimomys* and *Mimomys* whose remains compose over 84% of the bone assemblage. Prevalent in the Uryv-2 assemblage is the cement form *Mimomys* (1033 specimens, or 73.26%), whereas abundance rate of *Promimomys* is approximately 7 times lower (142 specimens, or 10.07%). Characteristic of the community was a low proportion of Lagomorpha forms (53 specimens, or 3.9%) and presence of forest dormouse, common field mouse, rat-like hamsters, and abundant insectivores (147 specimens, or over 10%). Among the latter, there were abundant and diverse shrews of the genus *Sorex*, larger *Blarinoides* and *Beremendia* forms, and abundant desmans.

The morphological analysis of voles from the Uryv-2 locality elucidated their evolutionary level. For instance, *Mimomys* ex gr. *polonicus* is a transitional form between *M. hajnackensis* from the Uryv-1 locality and *M. polonicus* from the Rebielice Krolewski locality (Kowalski, 1960a), being likely closer to the latter. *Mimomys* aff. *hintoni* is similar to the same species from the Seletino Formation of West Siberia (Zazhigin, 1980; Zazhigin and Zykin, 1984, Zykin *et al.*, 1987). By the molar characteristics, *M. pliocaenicus minor* from the Uryv-2 assemblage resembles this species from the Hajnacka and Beregovoi localities. On the other hand, it is more archaic than *M. pliocaenicus* described from the Norwich Krag site in Norfolk or from Shell Krag locality in East Runton and known as well from the Kadzielnia, Liventsovka, Podpusk, Lebyazh'e, and Kizikhi localities. *Promimomys gracilis* has more progressive features as compared with older representatives of the taxon. *Promimomys baschkirica* is also evolutionary more advanced than the species from the Uryv-1 locality. *Stachomys igrom* is larger and more similar to present-day *Prometheomys* than *Stachomys trilobodon* from the Weze locality of Poland (Kowalski, 1960b: Agadzhanyan, 1993).

As a whole, small mammals from the Uryv-2 locality belong to younger Uryv faunas and are best correlative with the Akkulaevo and Rebielice Krolewski faunas. The community composition is indicative of open forest–steppe landscapes with a significant proportion of broad-leaved forests at the thanatocoenosis burial time, and of near-water biotopes in warm climatic environments.

Bone Burials near the Korotoyak Village

Another group of burials is located on the right side of the Don River 10 km downstream of the Uryv Village, in the southeastern outskirts of the Korotoyak Village of the Ostrogozhsk district, Voronezh oblast. They were discovered in 1974 by Krasnenkov, who studied them (with mine participation) over 15 years (*Verkhnii pliotsen...*, 1985). In a system of ditches, there is exposed the following sequence (from the base upward): alluvial formations of the Don and Tikhaya Sosna (its tributary) rivers, ancient buried soils, Don Till, and, above the latter, loess and fossil soil horizons. The section and stratigraphic position of ditches have been described in detail earlier (Agadzhanyan and Glushankova, 1988, Iosifova and Semenov, 1998), and presented below is a brief characterization only. In ditches 1 and 2 near the ravine head, an upper part of the section is composed of the following beds:

(1) Recent soil (0–1.2 m thick);

(2) Loess-like loam, dark pale-yellow, porous, massive, enclosing carbonate concretions and pseudomy-

cellium, the bed has uneven lover boundary (1.2–1.7 m thick);

(3) Fossil soil I (1.7–2.2 m thick);

(4) Loess-like loam, dark pale–yellow with bluish tint, porous, massive, with abundant micellar carbonate material and rare incrustations of Fe and Mn hydroxides (2.2–3.55 m thick);

(5) Fossil soil II (3.55–4.25 m thick);

(6) Loam, grayish to dark pale–yellow, porous, calcareous, with slightly ferruginous spots and single films of Mn hydroxides, the lower boundary is uneven (4.25–4.75 m thick);

(7) Fossil soil III (4.75–6.30 m thick);

(8) Alternating grayish brown to grayish ocherous loam and fine-grained sand with ferruginate coloration slightly decreasing downward; the lower contact is gradual (6.30–7.55 m thick);

(9) Fossil soil IV (7.55–8.5 m thick);

(10) Sand, coarse-grained to inaequigranular, whitish yellow, ocherous, horizontally and cross-bedded, enclosing bed rock fragments, mud balls, gastropod shells, and bones of small mammals at depth of 8.9 m; the lower boundary of the bed representing locality Korotoyak-4 is even (8.5–9.05 m thick);

(11) Loam, variegated (dark ocherous, yellowish green, bluish gray, yellowish dark brown), with a horizon of highly weathered granite boulders and sandstone at depth of 10.0 m; rocks are highly ferruginous at the lower and upper contacts. Some intervals are enriched in carbonate material, and the interval from 12.24 to 14.5 m below the top of gleyed and highly ferruginous sediments contains rare pebbles of Cretaceous rocks; the loam (Don Till) has an even lower boundary (9.05–14.5 m thick);

(12) Intercalations of highly ferruginous dark-gray loam and whitish yellow fine-grained sand (14.5–15.65 m thick).

The section of morainal and underlying sediments is most complete in the middle part of the ravine (ditch 7), where the following beds are exposed (from the top downward):

(1) Recent soil (0.0–1.1 m thick);

(2) Loam and clay with an admixture of fine-grained sand, grayish brown to brown and dark brown with bluish tint; sand with rare boulders of local and crystalline rocks is calcareous and ferruginate (Don Till), the lower contact is even (1.1–4.8 m thick);

(3) Sand, fine-grained, ocherous–yellow, horizontally and cross-bedded, grading downward to alternating grayish yellow horizontally bedded sandy and loamy varieties; in the depth interval of 6.0–7.1 m (Korotoyak-3 locality), sediments bear rare shells of terrestrial gastropods and bones of small mammals (4.8–8.6 m thick).

Ditch 8 at the ravine mouth recovered the most complete basal part of the section. The following beds are distinguished here (from the top downward:

(1) Sand, light brown, horizontally bedded, with small pebbles of carbonate rocks, freshwater gastropod shells, and bones of small mammals (Korotoyak-3 locality, 0.0–5.8 m thick);

(2) Fossil soil I below the till corresponding to a humus horizon intensely colored by organic matter (5.8–6.0 m thick);

(3) Loam, bluish, with abundant pebbles of Cretaceous rocks and fragments of molluscan shells (6.0–6.7 m thick);

(4) Fossil soil II (6.7–7.2 m)

(5) Sand, fine-grained, light brown, horizontally bedded (7.2–10.4 m thick);

(6) Sandy loam and sand medium- to fine-grained, colored light brown, grayish yellow, gray, grayish green, horizontally bedded, bearing shells of gastropods and unionids, abundant well rounded pebbles of Cretaceous rocks, and enclosing lenses of cross-bedded sand; fragments of large and small mammals occur in the intervals of 10.8–11.3 and 11.3–11.6 m, the latter are most abundant in the last interval corresponding to locality Korotoyak-1 (10.4–12.2 m thick).

As is evident from the description, the basal part of the Korotoyak section is composed of two members of alluvial sands referred to the Korotoyak Formation and lower Uryv Subformation (Fig. 2) The Korotoyak Formation is crowned by floodplain loams with two fossil soils (Beds 2 and 4 in ditch 8). Both alluvial formations enclose remains of small mammals: the Korotoyak-1 and Korotoyak-2 bone-bearing horizons. They are overlain by the alluvial upper Uryv Subformation up to 3 m thick that yields remains of small mammals and corresponds to the Korotoyak-3 bone-bearing horizon. The lower part of the alluvium with normal polarity is correlated with the Réunion 1 Subchron (Iosifova and Semenov, 1998). The upper Uryv Subformation is overlain by the Tikhaya Sosna Formation, which also yielded fragments of Late Pliocene small mammals (the Korotoyak-2b bone-bearing horizon). The interval of Tikhaya Sosna Formation is of the reversed polarity (Iosifova and Semenov, 1998). In addition, the section includes younger Korotoyak-3 and Korotoyak-4 bone-bearing horizons, which contain the Taman and Tiraspol faunas. Figure 2 shows position of the bone-bearing horizons in the section.

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Insectivora	Abundance	%
Desmana cf. *senseyi* Kormos	5	0.35
Desmana nehringi Kormos	26	1.84
Talpa ex gr. *minor* Freudenthal	5	0.35
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(3) Fossil soil I (1.7–2.2 m thick);

(4) Loess-like loam, dark pale–yellow with bluish tint, porous, massive, with abundant micellar carbonate material and rare incrustations of Fe and Mn hydroxides (2.2–3.55 m thick);

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(6) Loam, grayish to dark pale–yellow, porous, calcareous, with slightly ferruginous spots and single films of Mn hydroxides, the lower boundary is uneven (4.25–4.75 m thick);

(7) Fossil soil III (4.75–6.30 m thick);

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(10) Sand, coarse-grained to inaequigranular, whitish yellow, ocherous, horizontally and cross-bedded, enclosing bed rock fragments, mud balls, gastropod shells, and bones of small mammals at depth of 8.9 m; the lower boundary of the bed representing locality Korotoyak-4 is even (8.5–9.05 m thick);

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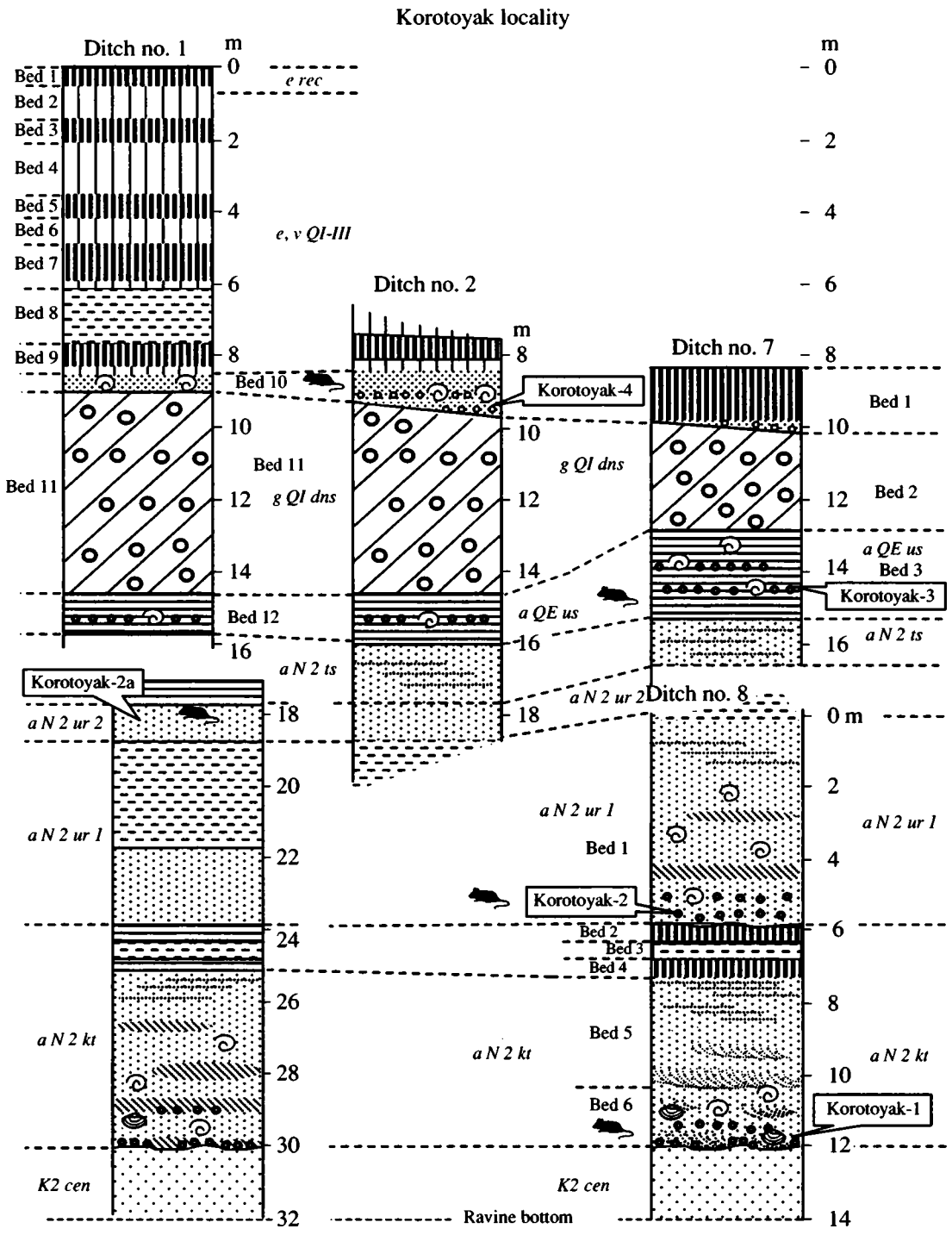

Fig. 2. Lithostratigraphy of Pliocene–Pleistocene sections near the Korotoyak Village, the Voronezh oblast (symbols as in Fig. 1; indices of beds after Krasnenkov *et al.*, 1987).

among rodents. Teeth of their dominant species *Promimomys gracilis* Kretzoi are of a low hypsodonty, having low tracks and deep position of marks on M^1 and M^3. *Promimomys baschkirica* Suchov is of archaic appearance. Primitive features are also characteristic of voles from the *Mimomys* ex gr. *hajnackensis* Fejfar lineage. These have lesser hypsodonty of molars, less developed dentine tracks, and marks later closing and longer preserved on M^1 and, particularly, on M^3, as compared with voles of this group, which are known from the Uryv-1, Korotoyak-2, Akkulaevo, and Rebielice Krolewski localities (Sukhov, 1970; Kowalski, 1960a).

In its morphology, relatively abundant *Pliomys ucrainicus* from the Korotoyak-1 locality is very similar to voles from the oldest population of *Pliomys ucrainicus topacevski* Nesin and to those from the middle Kotlovina Bed (Nesin, 1983).

The Korotoyak-1 assemblage of small mammals is best correlative with the fauna from the middle Kotlovina Bed of Ukraine and, probably, with the San-Giusto fauna of Italy (Topachevskii and Nesin, 1989; Masini and Torre, 1987).

In Table 4, there is a list of small mammal taxa from the Korotoyak-2 bone-bearing horizon. Voles determine the appearance of this fauna. Contrary to older Don River localities, proportion of hares in this assemblage is only 6%. Most abundant amid dominant species is the vole species *Promimomys gracilis* Kretzoi, which is morphologically close to its counterpart from the Korotoyak-1 and Uryv-1 assemblages. Teeth of *P. gracilis* from the Korotoyak-2 locality differ in a higher average height of the crown, stronger development of dentine tracks, and earlier disappearance of a mark on M^1. The distinctions are most contrast in comparison vole teeth from the Korotoyak-1 locality. In the development of hypsodonty and dentine tracks, larger *Mimomys hajnackensis*, other dominants of the assemblage, are close to the nominative population from the Hajnacka locality (Fejfar, 1961). This taxon is well correlative with *M. polonicus* from the Seletino Formation of the Irtysh River basin, being more advanced in evolution than *M. hajnackensis* from the Korotoyak-1 locality, but more archaic as compared with *M. polonicus* from the Arondelli locality (Zazhigin, 1980; Zazhigin and Zykin, 1984; Zykin *et al.*, 1987, Fejfar and Heinrich, 1983). *M. polonicus* populations from the Liventsovka, Rebielice Krolewski, and Perrier-Etouaires localities are more advanced (Aleksandrova, 1976; Kowalski, 1960; Chaline, 1974). Another vole group relatively abundant in the Korotoyak-2 locality includes representatives of the genus *Villaniya*. Small *Villaniya veterior* Kretzoi resembles much *Villaniya steclovi* Zazhigin from the Seletino Formation of West Siberia (Zazhigin, 1980), however, this taxon has more developed dentine tracks. *V. exilis* from the Liventsovka quarry is larger in dimensions than small *Villaniya* from the Korotoyak-2 locality. The less developed mimomys cusp on M^1 and better dentine tracks indicate a more advanced evolutionary level of this taxon. Gliridae of the Korotoyak-2 assemblage are unique. Their primitive teeth close in structure to those of *Dryomys* and *Myomimus* are larger than all known specimens belonging to these genera. Among insectivores there is shrew *Beremendia* sp. of a very large size. Thus, the majority of taxa from the Korotoyak-2 fauna are characteristic of the middle Pliocene. This assemblage is undoubtedly younger than the Korotoyak-1 fauna and small mammals from the lower Kotlovina beds and Seletino Formation of the Irtysh River basin. The Korotoyak fauna is close to faunas from the Uryv-1, Hajnacka, Arondelli, and other localities. Simultaneously, it is clearly older than small mammals from the Uryv-2, Rebielice Krolewski, Perrier-Etouaires, and other localities.

Table 3. Taxonomic composition of small mammals in the Korotoyak-1 locality

Insectivora	Abundance	%
Blarinoides sp.	2	0.41
Desmana thermalis Kormos	19	3.85
Talpa csarnotani Kretzoi	4	0.81
Erinaceus sp.	1	0.20
Lagomorpha		
Proochotona ex gr. *gigas* Arg. et Pidopl.	4	0.81
Ochotonoides danubicus Topacevski	3	0.61
Ochotonidae gen.	7	1.42
Pliolagus brachignatus Kormos	106	21.50
Rodentia		
Apodemus ex gr. *silvaticus* L.	2	0.41
Cricetulus (*Tcherskia*) sp.	20	4.06
Baranomys lozyci Kormos	3	0.61
Pliomys ucrainicus topacevski Nesin	41	8.32
Villanyia veterior Kretzoi	2	0.41
Promimomys gracilis Kretzoi	39	7.91
Promimomys baschkirica Suchov	17	3.45
Mimomys ex gr. *hajnackensis* Fejfar	6	1.22
Mimomys cf. *hintoni* Fejfar	12	2.43
Promimomys sp.– *Mimomys* sp.	171	34.69
Trogontherium sp.	16	3.25
Nannospalax odessanus Topacevski	18	3.65
Total	493	100.00

The third bone-bearing horizon (Korotoyak-2a) is confined to the upper member of the upper Uryv Subformation (Selyanovo sequence). Paleontological material from this locality was collected by Krasnenkov. The well-preserved bone remains include here the intact fragments of mandibles belonging to insectivores, voles, and others. Cement in entering corners of molars bears no signs of destruction, teeth roots are intact, and indications of transportation are invisible. The preservation degree implies that the taphocoenosis consists of materials subjected to a minimal reworking. The taxonomic composition of small mammals from Korotoyak-2a bone-bearing horizon is presented in Table 5. The assemblage includes abundant (up to 12%) remains of insectivores, among which there are identified small hedgehog, mole, shrews of the *Drepanosorex*, *Beremendia*, and *Blarinoides* genera, and small musk forms. Abundant are *Pliolagus brachygnathus*, and *Ochotona* forms (up to 10%). The assemblage includes also flying squirrel, dormouse, forest mouse, and *Stachomys igrom*, an endemic species of

Table 4. Taxonomic composition of small mammals in the Korotoyak-2 locality

Insectivora	Abundance	%
Beremendia sp.	2	1.36
Desmana thermalis Kormos	4	2.72
Talpa sp.	1	0.68
Lagomorpha		
Pliolagus sp.	16	10.88
Rodentia		
Myomimus sp.	3	2.04
Dryomys sp.	3	2.04
Apodemus sp.	2	1.36
Cricetulus (*Tcherskia*) sp.	20	13.61
Pliomys ucrainicus Topacevski et Scoiric	6	4.08
Villanyia veterior Kretzoi	9	6.12
Villanyia petenyii (Mehely)	14	9.52
Promimomys ex gr. *gracilis* Kretzoi	15	10.20
Mimomys hajnackensis Fejfar	12	8.16
Mimomys-Promimomys sp.	31	21.09
Nannospalax odessanus Topacevski	9	6.12
Total	147	100.00

the Don River basin. The main constituting forms of the Korotoyak-2a assemblage are ancient rhizodont voles, whose proportion is as high as 64%. Prevalent are low-cement and cement forms, whose quantity is three times higher than that of cement-free voles of the *Promimomys* genus. The Korotoyak-2a community undoubtedly belongs to later faunas of the Uryv assemblage. In the track height, voles of the *Villaniya* genus are more advanced as compared with their counterparts from the Uryv-2 locality, although the difference is insignificant. Representatives of the group *Promimomys (Cseria) baschkirica* Suchov are also slightly different: their M^1 disappears earlier in the ontogenesis, i.e., they are evolutionary more advanced as compared with voles from the Uryv-2 locality.

In the development level of voles, the Korotoyak-2a fauna is slightly older than the Khaprov faunal parastratotype, the lower sequence of the Liventsovka section. It is very close to the fauna from the Uryv-2 locality of the Don River basin and from the Rebielice Krolewski locality of Poland.

MIDDLE PLIOCENE SMALL MAMMALS FROM THE VOLGA RIVER BASIN

The Apastovo Locality

The nearest region with localities of the Uryv-type fauna is located in the Volga River middle courses. One of the localities is known here in the claydite quarry near the town of Apastovo in Tatarstan. O.E. Chumakov was first to collect remains of small mammals from this locality in 1987. Later on, I also sampled small mammal remains from this locality. The Apastovo quarry located on the right side of the Sviyaga River (Fig. 3) is 200 m across and over 20 m deep. All beds dip here from the southeast (from the watershed) to the northwest toward the Sviyaga River valley. The ditch A-1 in the northern quarry wall recovered the following beds (from the top downward):

(1) Variegated loam and clay with lenses and interbeds of gray compact loam and with clots and lenses of black clays and sands; recent quarry talus (0.0–0.3 m thick);

(2) Loam, dark gray to almost black, splintery, with ferruginous films along crack planes. The lower boundary is gradual; recent soil (0.3–0.7 m thick);

(3) Clay, gray, compact, splintery, becoming bleached downward. Whitish horizontal carbonate and rusty ferruginous laminae occur throughout the entire bed. Horizontal lamination in the lower part is emphasized by fine-grained sand laminae. The lower boundary is sharp (0.7–1.45 m thick);

(4) Sands fine-grained in the upper part turn into coarse-grained cross-bedded downward, they look variegated owing to alternating light brown and gray laminae. The latter contain abundant small-sized plant remains. Throughout the bed there are scattered shell debris of freshwater gastropods. Sands from the lower part of the bed contain gastropod and large unionid shells, ferruginous concretions, and small wood fragments. The basal part of the bed yields bones of fishes and large and small mammals. The lower boundary is undulating, very sharp, and ferruginate. The Kumurlin Horizon, N_2a_1 km (1.45–2.5 m thick);

(5) Clay, compact, gray with bluish tint, enclosing rare horizons of ferruginate concretions (1–5 cm across) and scattered shells of small freshwater gastropods. The lower boundary is gradual, laterally well sustained. The Cheben'kovo II-III Horizon, $N_2 k_{1-3}$ tsh II–III (2.5–4.15 m thick);

(6) Clay, gray with brownish tint, splintery, with slight horizontal bedding, enclosing several horizons of ferruginate concretions 10–15 cm across. The lower boundary is gradual (4.15–5.2 m thick);

(7) Silt and clay, dark gray to greenish (tobacco), locally showing thin horizontal lamination. Sediments resemble foliated marl. Some laminae enclose plant remains and shells of small gastropods. The lower boundary is gradual (2.5–6.3 m thick);

(8) Clay, dark gray with brownish tint, compact, with slight horizontal lamination and well-developed ferruginous films along crack planes. The lower boundary is gradual (6.3–6.7 m thick);

(9) Clay, dark gray to almost black, moist, with thin horizontal lamination seen in dry sediment. The exposed interval is 6.7–7.0 m.

Sands of Bed 4 referred to the Kumurlin Horizon (Yakhimovich *et al.*, 1997) represent the main bone-bearing level. They rest upon the partly eroded main sequence of compact gray clays with the facies appearance varying through the section. The depth to the base of alluvial formation that corresponds to bone-bearing horizon ranges from 1.5 m in the southeastern part of the quarry to 6–7 m in its northwestern part. The thickness of the bone-bearing horizon proper varies from 0.2 to 0.6 m. In ditches A-1 and A-2 located closer to the watershed (southeastern wall), the horizon is composed of less coarse, silty material, which contains rare wood fragments and large bivalve shells, is also rich in shells of small gastropods. Bones of small mammals are relatively well preserved here, although remains of larger animals, e.g., of hares and beavers, are rare. According to Yakhimovich *et al.* (1997), the Kumurlin sands in the Apastovo section, which show in general the normal polarity, include an interval of the reversed polarity in the lower part of the sequence. This interval is correlated with the Kaena Subchron of the Gauss Chron.

Bone remains from the basal part of Bed 4 are varicolored. They are mostly dark brown, although some of them are colored noticeably lighter. Some teeth bear signs of reworking and digesting tract impact. All the remains are fossilized. Approximately 500 remains of small mammals and several tens of fish bones were examined. The taxonomic composition of the assemblage from the Apastovo bone-bearing locality is presented in Table 6. According to E.K. Sychevskaya who determined fish taxa, the Apastovo fauna includes remains of pike, Black Sea roach, tench, pike perch, roach, and crucian carp in addition to forms listed in Table 6. The ichthyofauna composition indicates its habitat in the lacustrine or lacustrine–fluvial environments and temperate climatic conditions.

Insectivores played a noticeable role (3.8%) in the mammal assemblage under consideration. These were large Pliocene shrews of the genus *Blarinoides*, moles, and desmans. Hare species *Pliolagus brachygnathus* (14.25%) and rhizodont voles (78.2%) in association with beavers and rhizodont zokor of the genus *Prosiphnaeus* were abundant. According to the community composition, climatic conditions were quite favorable in its habitat.

The composition and evolutionary level of main groups suggest a concrete age for the Apastovo fauna. Abundant lagomorphs (over 14%), rhizodont zokor, *Prosiphnaeus*, and almost all mentioned vole species are characteristic of the middle Pliocene, namely of the early Villafranchian faunas. The same age is indicated by occurrence of shrew *Blarinoides*, small mole, and others. Vole *Mimomys* ex gr. *hajnackensis* Fejfar is an advanced form of this species populations. Its molars have dentine tracks developed somewhat better than in teeth of the species from the Hajnacka type locality of Czechia (Fejfar, 1961b, 1964). *Promimomys* and *Villaniya* forms are very close to voles of both genera from the Uryv-1 and Korotoyak-2 localities.

Table 5. Taxonomic composition of small mammals in the Korotoyak-2a locality

Insectivora	Abundance	%
Erinaceus sp.	1	0.18
Talpa sp.	4	0.73
Desmana aff. *termalis* Kormos	18	3.29
Petenyia hungarica Kormos	6	1.10
Beremendia fissidents (Petenyi)	10	1.83
Blarinoides marinae Sulimski	12	2.19
Drepanosorex sp.	2	0.37
Soricini indet.	12	2.19
Lagomorpha		
Ochotona sp.	3	0.55
Pliolagus brachignatus (Kormos)	52	9.51
Rodentia		
Pliopetaurista sp.	1	0.18
Myomimus sp.	1	0.18
Apodemus ex gr. *silvaticus* L.	3	0.55
Apodemus sp.	2	0.37
Allocricetus sp.	19	3.47
Cricetulus sp.	6	1.10
Stachomys igrom Agadjanian	19	3.47
Villanyia exilis Kretzoi	3	0.55
Promimomys (Cseria) baschkirica Suchov	32	5.85
Mimomys altenburgensis Rabeder	20	3.66
Mimomys polonicus Kowalski	47	8.59
Mimomys hintoni Fejfar	16	2.93
Promimomys aut Mimomys	213	38.94
Nannospalax odessanus Topacevski	40	7.31
Castoridae gen. indet.	3	0.55
Carnivora		
Mustela sp.	2	0.37
Total	547	100.00

THE URYV FAUNAL ASSEMBLAGE

Thus, the described sections demonstrate a succession of sediments and bone-bearing horizons, which characterize evolution of small mammal communities belonging to a common faunal assemblage. In taxonomic composition and evolutionary level of taxa, the assemblage is transitional between the Moldavian and Khaprov faunas. Its characteristic features are as follows:

(1) Prevalence of rhizodont voles of the *Promimomys–Mimomys* group;

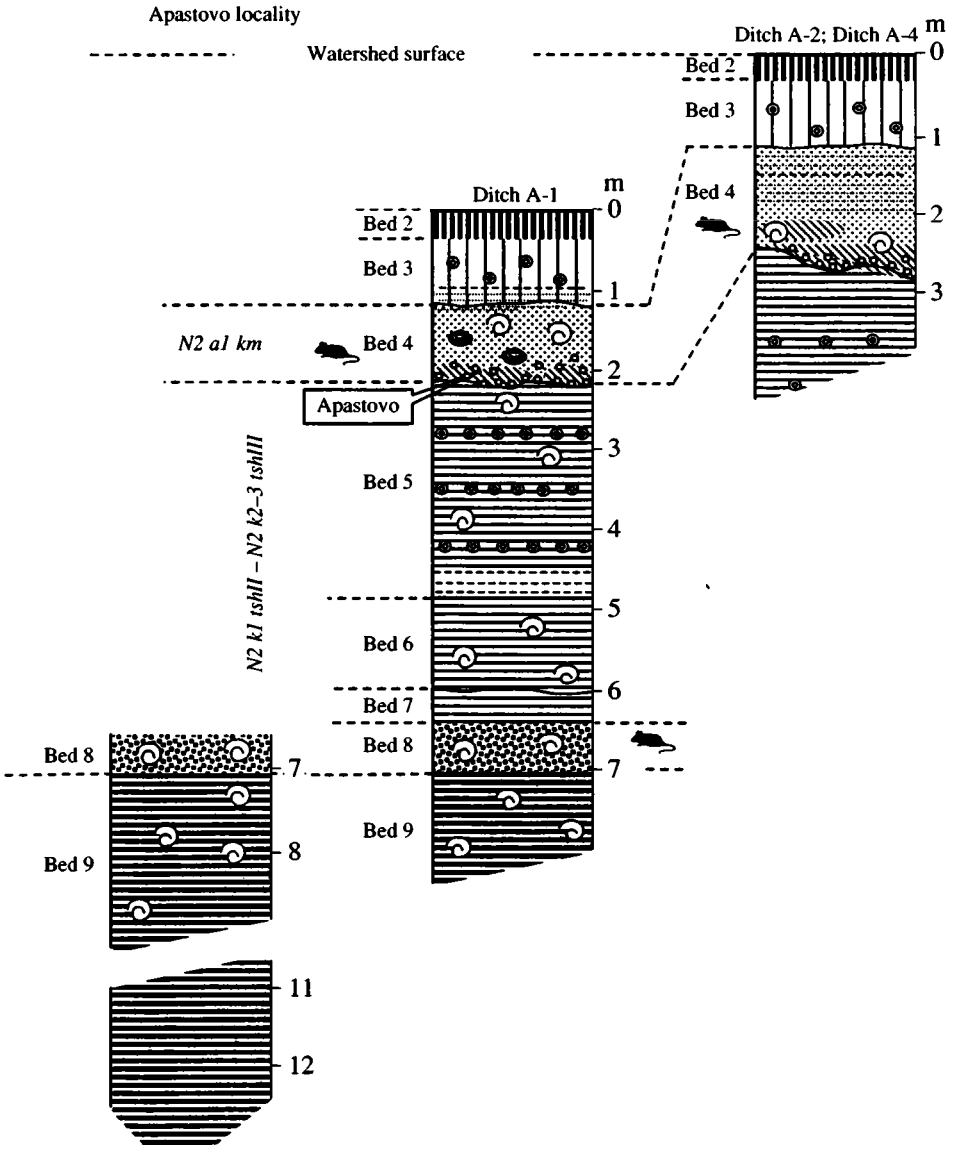

Fig. 3. Lithostratigraphy of Pliocene sections near the town of Apastovo, Tatarstan (symbols as in Fig. 1; indices of beds after Yakhimovich *et al.*, 1997).

(2) Relatively high abundance (10 to 20% in average) of lagomorphs;

(3) Presence of *P. ucrainicus* vole group of the genus *Pliomys*;

(4) Absence or extreme scarcity of voles of the genus *Dolomys*;

(5) Common occurrence, sometimes noticeable, of primitive voles from the genus *Villaniya*, e.g., of *V. exilis*;

(6) Presence of rare vole-like rodent of the genus *Baranomys*;

(7) Occurrence of peculiar primitive voles belonging to the genus *Stachomys*, the ancestral taxon of recent *Prometheomys* forms dwelling in the Caucasus;

(8) Taxa most characteristic of different stages in fauna evolution were voles *Promimomys gracilis* Kretzoi, *Promimomys baschkirica* Suchov, *Mimomys hajnackensis* Fejfar, and *Mimomys polonicus* Kowalski;

(9) Insectivores are represented by archaic shrews of the *Blarinoides, Beremendia*, and *Drepanosorex* genera and by small Pliocene desmans of the genus *Desmana*.

In terms of the Land Mammal Ages, the Uryv assemblage corresponds to the first half of the Villanian, i.e., to the lower Villafranchian. Its early and advanced communities are correlative with small mammals of Subzone MN16a and the latter faunas correspond to those of Subzone MN16b.

The alluvial sediments at the base of the Pliocene–Pleistocene sequence exposed near the Uryv Village, the Ostrogozhsk district of Voronezh oblast (Uryv-1 bone-bearing horizon, or locality) can be considered as representing a stratotype for the Uryv faunal assemblage. These sediments rest upon Cretaceous sands. Upward the section, they grade into floodplain loam with signs of subsequent soil formation. In turn, the erosional surface of loam is overlain by alluvial–lacustrine clays of the Uryv-2 locality.

The Uryv faunas are easily distinguishable from older communities of small mammals referred to the Moldavian faunal assemblage. For instance, the proportion of lagomorphs exceeds 20% in the older Gerasimovka fauna from the Oskol River basin and 30% in even older Antipovka and Chugunovka faunas from the Don River basin. In the Gerasimovka fauna, voles of the *Promimomys–Mimomys* group (slightly over 15%) are represented by most primitive *Promimomys moldavicus* Kormos. Dominant among voles are representatives of the *Dolomys* and *Pliomys* genera (22 and 23%, respectively). The Moldavian faunas occupy quite definite position in the stratigraphic scale. The lower bone-bearing level of the Kotlovina locality and the middle horizon of the Etuliya locality are confined to upper intervals of the Gilbert Chron (Vangengeim *et al.*, 1995, 1998). According to the morphometric analysis, vole populations from the lower and middle horizons of the Etuliya locality are 3.61 and 3.583 Ma old, respectively (Pevzner and Vangengeim, 1994). Inasmuch as communities of the Moldavian assemblage are evolutionary more archaic than their Uryv counterparts, it is safe to suggest that the latter are younger than the Gilbert/Gauss boundary time.

The Uryv faunas are sufficiently well distinguishable also from the later Khaprov communities. Lagomorphs are noticeably less abundant (3–5%) in the latter, which simultaneously show the higher abundance and diversity of voles representing the genus *Mimomys*. In addition, characteristic of the Khaprov communities are the lower abundance and diversity of *Promimomys* forms, scarcity of taxa from the *Pliomys ucrainicus* group, and almost complete absence of *Dolomys* species. Voles of the genus *Villaniya* are represented in these communities by evolutionary advanced forms with teeth bearing the well-developed dentine tracks.

As is evident from the above descriptions, the Uryv communities represent a distinct succession reflecting the faunal evolutionary stages. As one can judge from the Korotoyk-1 locality, lagomorphs were still abundant (approximately 20%) in the early (in terminology of I.M. Gromov) Uryv community, in which voles of the genus *Promimomys* were much more abundant (approximately two times) than *Mimomys* forms of a low species diversity whose teeth had a low content of cement. In addition, proportion of *Pliomys* species in this community was as high as 8%.

Table 6. Taxonomic composition of small mammals in the Apastovo locality

Insectivora	Abundance	%
Desmana sp.	11	2.41
Talpa ex gr. *minor* Freudenthal	2	0.44
Blarinoides sp.	4	0.88
Lagomorpha		
Ochotona sp.	3	0.66
Pliolagus brachignatus Kormos	64	14.00
Rodentia		
Tamias orlovi Sulimski	3	0.66
Villanyia veterior Kretzoi	7	1.53
Promimomys (*Cseria*) *baschkirica* Suchov	26	5.69
Promimomys sp.	23	5.03
Mimomys cf. *hintoni* Fejfar	20	4.38
Mimomys ex gr. *hajnackensis* Fejfar	25	5.47
Promimomys-Mimomys	250	54.70
Lemmini gen.	2	0.44
Prosiphnaeus sp.	15	3.28
Castoridae gen.	2	0.44
Total	457	100.00

In the advanced Uryv faunas, percentage of lagomorphs is lower (approximately 8–12%), whereas *Promimomys* and *Mimomys* forms are present in comparable proportions, and the latter are more specialized. *Pliomys* species are rare (4%) in these faunas. In contrast, vole *Mimomys hajnackensis* Fejfar, the index species of Subzone MN 16a (Fejfar *et al.*, 1998), is rather abundant like in older communities.

In the late Uryv faunas, the vole *Mimomys* genus substantially prevails over *Promimomys* species whose abundance is 2–3 times lower. The proportion of lagomorphs decreases up to 3–4% in these faunas. The species diversity and evolutionary level of voles noticeably increases: their crowns and tracks of buccal teeth are higher, and amount of external cement is greater. The important role in these communities belongs to species *Mimomys polonicus* Kowalski characteristic of Subzone MN 16b (Fejfar *et al.*, 1998).

The morphology of main mammal taxa from Zone MN 16 of Central and West Europe is described in detail (Bachelet, 1990; Bruijn *et al.*, 1992; Esteban and Lopez Martinez, 1990; Fejfar, 2001; Fejfar and Heinrich, 1982, 1990; Masini and Torre, 1987, 1990; Rabeder, 1981; Döpps and Rabeder, 1997; Storch and Fejfar, 1990). These materials show that small mammals of the Uryv assemblage are well correlative with those from the European localities Moreda, Sete, Balary-2, Arondelli, Hajnacka, Beremend-5, Rebielice Krolewski-1, Deutsch-Altenburg-20, Deutsch-Alenburg-21, Stran-

Table 7. Position of Uryv faunas in the Pliocene–Pleistocene succession of small mammal communities in the Russian Plain

System	Series	Subseries	Stage	Ma	Chronological boundaries	Magneto-stratigraphy	Regional stage	MN and MQ zones	Russian Plain: Mammal assemblages	Localities in the Don and Dniester river basins	Main small mammal species
Quaternary	Pleistocene	u. middle	Emilian	0.5	0.78	Brunhes		MQ21	Paleolithic	Mikhailovka-1	*Arvicola* cf. *sapidus*, Microtus arvalis
									Khazar	Topka	*Arvicola chosaricus*
									Singul	Vladimirovka, Strelitsa	*Arvicola mosbachensis*, *Lagurus transiens*, *Microtus malei*
								MQ20	Tiraspol	Kuznetsovka (+)	*Mimomys intermedius*, *Lagurus posterius*, *Terricola gregaloides*
										Novokhopersk, Veret'e (+), Il'inka	*Mimomys intermedius*, *Mimomys pusillus*, *Microtus oeconomus*, *Terricola hintoni*
		lower	Calabrian	1		Kz, J, K-M	Apsheronian	MQ19	Taman	Moiseevo-1 (+), Log Krasnyi (–), Korotoyak-3c (+), Korotoyak-3b (–)	*Allophaiomys pliocaenicus*, *Terricola*, *Prolagurus pannonicus*, *Mimomys pusillus*
				1.5				MQ18	Odessa	Korotoyak-3a (–), Uspenka, Log Denisov	*Allophaiomys pliocaenicus*, *Prolagurus praepannonicus*, *Mimomys pusillus*
										Strelitsa-1 (–)	*Mimomys savini*, *Mimomys pliocaenicus*, *Allophaiomys pliocaenicus*
					1.8	Old				Mikhailovka-1	*Clethrionomys sokolovi*
Neogene	Pliocene	upper	Gelasian	2		R	Akchagylian	MN17	Khaprov	Liventsovka 4–5 (–), Liventsovka 1–3 (–), Krivskii (–)	*Mimomys pliocaenicus*
				2.5		Matuyama		MN16b	Uryv, late	Korotoyak-2a (–), Uryv-2 (–)	*Promimomys baschkirica*, *Mimomys polonicus*, *Mimomys pliocaenicus minor*, *Stachomys igrom*
		middle	Piacenzian		2.59				Uryv, advanced	Uryv-1 (–), Korotoyak-2 (+)	*Mimomys hajnackensis*, *Villanyia petenyii*, *Nannospalax odessanus*
				3		K, M		MN16a	Uryv, early	Korotoyak-1 (–)	*Pliomys ucrainicus topacevski*, *Promimomys gracilis*, *Mimomys* ex gr. *hajnackensis*
				3.5	3.58	Gauss		MN15	Moldavia	Korotoyak-Don (–), Gerasimovka,	*Dolomys nehringi*, *Pliomys jalpugensis*, *Promimomys moldavicus*
		upper	Zanclean	4		Co	Cimmerian	MN14	Kuchurgany	Antipovka (–), Chugunovka	*Pliopetaurista*, *Epimeriones*, *Polonomys insuliferus*, *Nannospalax macoveii*, *Trogontherium* ex gr. *minus*
				4.5		Nv					
						Sd					
				5		Tv					

Note: The table is compiled using materials by Yu.I. Iosifova on the Pliocene and Eopleistocene regional stages of the Upper Don basin approved by the Russian Interdepartmental Stratigraphic Committee for the central and southern Russian Plain, resolution of February, 2000. + and – in brackets designate normal and reversed polarity, respectively.

zendorf A-C, and others. Faunas from all these localities are not strictly synchronous. They reflect development of small mammals within the time interval corresponding to Zone MN 16 (Fejfar *et al.*, 1998). The oldest faunas are known from the Sete and Balaru-2 localities, and the youngest one is from the Rebielice Krolewski locality (Aleksandrova, 1989).

The data on the Akkulaevo and Simbugino localities in Bashkortostan (Sukhov, 1977) and my own study of bones from the Apastovo, Karan-Azikovo, Podgornye Bailary, and Deukovo localities in the Volga River basin show that the Uryv-type communities were of a wide geographic range. In addition to the Don River basin, they populated the Middle Volga region and Bashkortostan, showing however some regional and zoogeographic distinctions. For instance, mole-rat *Nannospalax odessanus* Topacevski was characteristic of the Don River communities, whereas rhizodont zokor *Prosiphnaeus* was typical of faunas in the Volga River basin, where first Lemmini also appeared at that time. The latter have not been found so far in the Don River basin. The *Villaniya* forms were abundant in southern areas, whereas in the Don upper reaches and Bashkortostan, they are scarce. On the other hand, dormouse and chipmunk dwelt in two latter areas being absent in the Black Sea region.

The age of the Uryv faunas is within the interval of 3.3 to 2.2 Ma, i.e., it is bracketed approximately between the Mammoth Subchron of the Gauss Chron and the Réunion Subchron of the Matuyama Chron (Table 7). The older and advanced Uryv faunas correspond to the second half of the Gauss Chron like in southern France, where the bone-bearing horizon of the Perrier-Etouaires locality is dated back to 2.6 Ma, i.e., corresponds to the end of the Gauss Chron. *"Mimomys pliocaenicus polonicus"* is a guide fossil of this locality (Chaline, 1996). The younger Uryv faunas correspond to the first half of the Matuyama Chron, and this geochronological interpretation is shared by many specialists (Aguirre *et al.*, 1996; Fejfar and Heinrich, 1982, 1983, 1987; Fejfar *et al.*, 1998). In West Siberia, the Kyzyl-Aigir fauna representing a close evolutionary analogue of the Uryv assemblage corresponds to the middle and upper parts of the Gauss Chron (Zykina *et al.*, 1995).

The Khaprov faunal assemblage, which corresponds to Zone MN 17, is terminal one in the Pliocene evolution of small mammals. In the southern Russian Plain, these small mammals are representative and have been described in detail (Shevchenko, 1965; Aleksandrova, 1967; Agadzhanyan *et al.*, 1976). In West Europe, they are known from the classical Saint Vallier locality, where bone-bearing sediments are 11 m thick in total, show negative polarity, and are correlative with the Matuyama Chron. According to data obtained by the magnetic resonance method, bone remains from the last site are approximately 2 Ma old (Valli, 2001). Accordingly, this fauna is close in age to the Liventsovka assemblage and younger than the Uryv fauna. The latter is thus older than 2 Ma.

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On Geochronological Significance of the Rb–Sr Study of Proterozoic Shales from the Anabar Massif

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Recently, Gorokhov *et al.*, (2001) published the paper, where they presented the results of the Rb-Sr system study in Upper Precambrian clayey rocks from the western slope of the Anabar Massif. The work is based on a perfect experimental material, but its interpretation seems insufficiently correct. Below, we consider some statements mentioned in this work.

The authors analyzed two shale samples from the Yusmastakh Formation, the upper one in the regional Riphean succession. The Yusmastakh Formation approximately 800 m thick is composed almost entirely of dolomites with rare thin (tens of centimeters) intercalations of terrigenous, largely clayey rocks. The latter varieties are confined to the basal parts of the upper and lower subformations, the latter resting on the erosional surface of underlying carbonate deposits. Both examined samples were collected precisely from these basal layers. The Rb–Sr measurements were carried out in clay subfractions, which were extracted from shales and subjected to the special preliminary treatment. The results obtained are interpreted by Gorokhov and his co-authors from the isochron standpoint, although the used methodology seems insufficiently substantiated. We try to demonstrate this based on the data from the mentioned paper by Gorokhov *et al.* (2001) and from their earlier work dedicated to the Ust'-Il'ya Formation of the Anabar Massif Proterozoic cover (Gorokhov *et al.*, 1997).

PECULIARITIES OF Rb–Sr DATING OF SEDIMENTARY ROCKS

As is known, the Rb–Sr dating method is applicable for the rock samples, which have the same initial isotopic ratio $(^{87}Sr/^{86}Sr)_0$. Mechanisms responsible for the equilibration are not always clear, but their function in natural settings is doubtless.

The mechanism of Sr isotope equilibration in the fine-grained fraction of sediments in the course of authigenic formation of clay minerals or during their exchange with the seawater Sr, the isotopic composition of which is constant during sedimentation, is a most simple explanation. The idea of using clay fractions to determine sedimentation time was put forward by M. Bonhomme with colleagues in 1966 (see in Clauer, 1984). This approach accepted by many researchers stimulated many works, which seem very promising for a long time. For instance, Clauer, an outstanding specialist in isotopic studies of clay fractions, was an apologist of this method. Subsequently, he rejected the idea, and we cite here a remarkable, from our viewpoint, sentence from one his work: "The K–Ar and Rb–Sr isotopic systems of sedimentary rocks and their mineral fractions have been used in many studies to determine the deposition time, although this approach is disputable. Despite an ambiguity of their application in stratigraphic research, the K–Ar and Rb–Sr systems are proven to be especially useful for determining the time and, in some cases, the duration of diagenetic processes" (Clauer *et al.*, 1992, p. 240).

Another mechanism of Sr isotope composition equilibration in sediments is an intense mixing of sedimentary material in the course of transportation and accumulation. In this case, the averaging may be attained within a certain volume of sediment (whole-rock sample) rather than in all or most grain-size fractions of the rock. Cordani *et al.* (1978) were among most active adherents of this approach. Studying clayey rocks from stratigraphic interval spanning the Upper Precambrian to Jurassic, they showed a good correspondence between the measured and true stratigraphic rock ages. However, it appeared eventually that such a correspondence is likely an exception rather than a rule in this case as well.

Finally, the third mechanism of Sr isotope equilibration in sedimentary rocks is their epigenetic transformation. Many details of this mechanism need in additional special studies, although the possibility of isotopic dating of epigenetic stages is beyond a doubt. Clauer with co-authors accentuate precisely this fact is in the above-mentioned sentence.

In the isochron diagrams, data points that meet requirements of isotopic dating plot along a straight line. In order to obtain the isochron dependence, two conditions are necessary in addition to the $^{87}Sr/^{86}Sr$ initial ratio equilibration in the samples: (1) the system

should be closed with respect of Rb and Sr and (2) the objects under examination should have different Rb/Sr values. All these requirements are well known and mentioned here with the only purpose to recollect once more that one cannot expect a confident Rb–Sr age determination for an event based on a single sample. At least two samples define an isochron, whereas three samples at minimum are necessary for obtaining the unambiguous result. The samples should differ in the Rb/Sr values, and the greater is the difference the lower are errors of calculated isochron parameters: the age and the initial Sr isotopic ratio. The issue is complicated by the fact that clay fractions of equal grain size from the same horizon are usually characterized by close Rb/Sr values, being hardly suitable for age determination.

ON INTERNAL ISOCHRONS

Meanwhile, there is an approach, which seemingly solves the problem of obtaining the required data scattering in the isochron diagram. The matter is that, in addition to Sr incorporated into a crystalline lattice of phyllosilicates, this element can be adsorbed to the surface of clay particles or can occupy some other positions not connected with the mineral crystalline structure. In the mica (muscovite, illite) structure, Sr may replace K or Rb, but such a replacement decreases the structure stability, because the elements are of a different valence and ion radius.

The easily mobile (exchangeable) Sr can be extracted from a sample using a weak solvent such as ammonium acetate (Gorokhov *et al.*, 1997, 2001) or 1N HCl solution (in many laboratories abroad). Rb replacing K in micas is significantly less soluble, and $^{87}Rb/^{86}Sr$ values appear therefore to be lower in leachates. Correspondingly, this ratio in residues after leaching is higher than in the untreated sample. Thus, two additional points, instead of one corresponding to an untreated sample, appear in the isochron diagram. In accord with the mass balance, these three points must plot along a straight line, of course provided a perfect quality of experimental equipment. As the data point corresponding to the leachate plots close to the origin of coordinates in the diagram, the errors in calculated parameters of the plotted straight line appear to be minimal. Taken together, this procedure creates an illusion of an accurate age determination.

The approach based on the preliminary leaching was repeatedly considered in publications. Sometimes, it leads to a reliable (even stratigraphically significant) age determination, but in most cases the results obtained cannot be unambiguously interpreted (Clauer *et al.*, 1990; Ohr *et al.*, 1991; Vinogradov, 1997), because the Sr easily extractable from a sample may have and sometimes actually has nothing to do with the Sr strongly bonded in the crystalline lattice of a mineral subjected to leaching. It is simply absurd from standpoints of lithology, geochemistry, and hydrogeology to assume that, being slightly bonded, this element may reside in rocks during many hundreds million years and characterizes environments of sedimentation or early diagenesis.

As a result, the skillful but artificial approach leads to a similarly artificial method of age calculation, although the calculated dates seem to be sufficiently plausible in many cases, and this is quite explainable. The $^{87}Rb/^{86}Sr$ values in clay fractions are usually relatively high. In concrete samples and leachates studied by Gorokhov *et al.* (2001), they range from 14 to 70 and from 2 to 10, respectively. Thus, the points, which correspond to leachates, plot closer to the origin of coordinates in the isochron diagram than the points corresponding to clay fractions. Therefore, the relatively small variations of $^{87}Rb/^{86}Sr$ values in leachates insignificantly change the straight line slope (and thus the calculated age) that is determined by points remote from the origin of coordinates and guarantee, instead, an insignificant error of the calculated tangent value. As a result, the estimated age can be approaching some real value, but determined with the unreal, artificially lowered errors of calculation. The age calculation errors almost do not depend on the Sr isotopic ratios measured in leachates. Par example, if we change artificially the $^{87}Rb/^{86}Sr$ value measured in a leachate from some fraction by a such value as ±0.05 (that is significant, exceeding by 100 times the analytical measurement error), the calculation error for the internal (three-point) rectilinear dependence will be so small that the dependence will be still isochronous. At the same time, the calculated age changes by ±50 ... ±100 Ma on the average.

Naturally, one can most closely approach the true age of an event taking into account only the data on residues after leaching. They represent samples free of highly mobile Sr probably alien to the mineral system proper. Keeping in mind the aforesaid, let us consider the factual data presented in the paper under consideration (Gorokhov *et al.*, 2001).

ISOCHRON PLOTS

The use of data on leachates is the most vulnerable point of the internal isochrons, although without them the latter loss their sense at all. One can try to make the situation easier uniting in calculations the data on different clay fractions, which also form frequently the linear trends in the isochron diagrams. Such procedure is performed for Fig. 10 in the paper under consideration to substantiate the idea that linear dependencies in the figure represent mixing lines. The weakness of the suggested substantiation is likely evident to the authors themselves, as they accompany it by the following phrase: "Thus, we can conclude that recent or slightly older interstitial solutions cannot be accepted for the

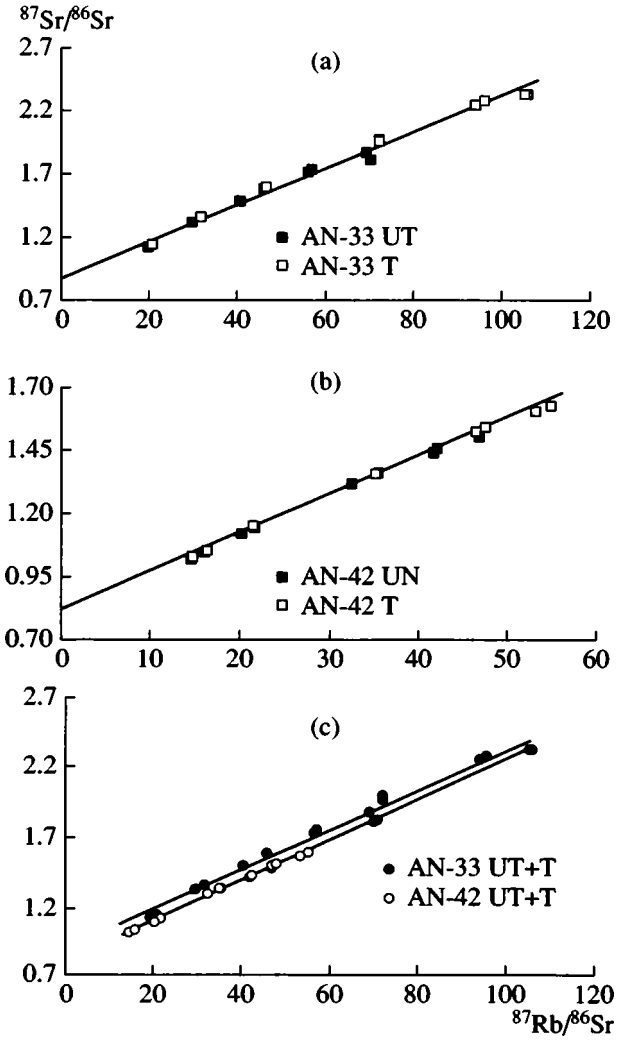

Fig. 1. Isochron dependencies for untreated (UT) and treated (T) size-variable clay fractions from two rock samples of the Yusmastakh Formation (after Gorokhov *et al.*, 2001): (a, b) separate plots for each sample; (c) generalized plot for two samples.

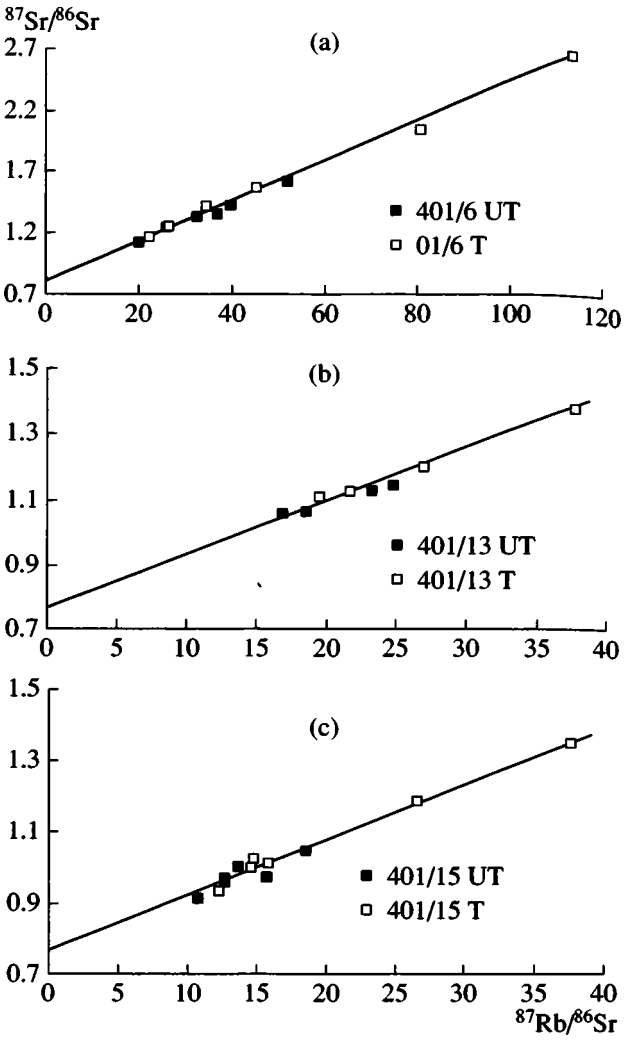

Fig. 2. Isochron dependencies for untreated (UT) and treated (T) size-variable clay fractions from three rock samples of the Ust'-Il'ya Formation (after Gorokhov *et al.*, 1997).

general source[1] of mobile strontium..." (Gorokhov *et al.*, 2001, p. 17).

Figures 1 and 2 in our paper reproduce plots for all grain-size fractions of lower and upper Yusmastakh shales, while the results of corresponding isochron calculations are presented in the table. A rather large scatter of points relative to approximating straight lines and lack of basic data points (leachates) near the origin of coordinates lead to large errors in parameters calculated for the lines. It is also remarkable that points corresponding to both the treated and untreated samples plot along the same line (Fig. 1a, b). To explain such a coincidence, one may assume that a minor percentage of substance entering the leachate does not disturb the substance balance in a given sample.

[1] Italicization by Gorokhov *et al.* (2001).

The table and Fig. 1 clearly demonstrate that all grain-size fractions of both samples collected from different subformations have similar ages and slightly different initial isotopic ratios, although a special strict statistical calculation shows that parameters of corresponding straight lines are identical. The age differences established by the authors (Gorokhov *et al.*, 2001) for clay subfractions of the samples is an artifact, the consequence of an incorrect idea that the "mobile" components and those bonded in minerals represent a single Rb–Sr isotopic system. It is possible to assume therefore that their "incomplete unity" exists in fact. The last assumption is substantiated in the following section.

NATURE OF MOBILE STRONTIUM

As is noted in the paper under consideration, "As the size range of subfractions ... decreases ... In leachates,

concentrations of Rb and Sr, as well as the $^{87}Sr/^{86}Sr$ ratios increase" (Gorokhov *et al.*, 2001, p. 11). Maybe it was a partial dissolution of the clay minerals proper responsible for that increase? In this case, exactly the matter of finest subfractions should be extracted by solvent, and precisely this process could be responsible for the remarkably high initial Sr isotopic ratios in shales. In addition, it should be taken into account that the studied clayey rocks have been sampled from thin interbeds in a continuous carbonate succession up to 800 m thick. The Sr isotope composition in the latter is extremely low (Pokrovskii and Vinogradov, 1991). The sharp contrast between the Sr isotope compositions in host rocks and clayey interbeds can be easier explained by predominant extraction of "mobile" Sr from their "immobile" mineral components. It is logical to assume that the leachates contained the mixed Sr: that absorbed by clay minerals and that extracted from the minerals proper. Moreover, in the considered case, the latter seems to be sharply prevailing over the former. Previously, it was shown that the mineral Sr is extractable from clay fractions in the 2.5 N HCl solution (Clauer, 1984). Precisely this stimulated the use of weaker solvents, a complete indifference of which relative to the mineral Sr is only assumed.

Proportions between different forms of extracted Sr depend on the mineral composition of examined fractions. In smectite, Sr and Rb are in exchangeable position both, whereas in the mixed-layered 2 : 1 minerals of the smectite-mica type, Rb is almost entirely incorporated into the mineral structure and all Sr cations are in exchangeable positions. In 2 : 1 micas proper, Rb is entirely concentrated between the layers, whereas Sr occupies this position only to a minor extent. By a diminishment of particle size, the sample treatment even in the ammonium acetate will unavoidably result in an increase of analytically significant exchange between the ammonium ion, on the one hand, and both the Sr and Rb bonded in mineral structures, on the other (Brown, 1965; Godovikov, 1983).

Previously, Gorokhov *et al.* (1997) published the results of a similar study of clayey rocks from the Ust'-Il'ya Formation, the basal one in the carbonate Billyakh Group of the Kotuikan section. Using the same method of internal isochrons (leachate–residue-untreated sample), they concluded that the rocks experienced two transformation stages (at approximately 1400 and 1190 Ma ago) and substantiated their inference assuming presence of two illite generations. Figure 2 in our paper demonstrates trends of data points corresponding to all clay subfractions, treated and untreated in the solvent, from three samples of the Ust'-Il'ya Formation examined by Gorokhov *et al.* (1997), and the results of relevant isochron calculations presented in the table. The point scattering and parametric errors of the straight line characterizing trend of the Ust'-Il'ya Formation samples are substantially high in accord with mainly terrigenous coarse-grained composition of the sampled rocks. Nevertheless, the age calculated for all samples is similar to that obtained for samples from the Yusmastakh Formation.

Results of isochron calculations (York, 1966) for clay fractions of treated (T) and untreated (UT) samples from the Yusmastakh and Ust'-Il'ya formations (Gorokhov *et al.*, 1997; 2001)

Sample	Number of measurements	Age, Ma	$(^{87}Sr/^{86}Sr)_0$
Yusmastakh Formation			
AN-33 T	12	1062 ± 70	0.845 ± 0.037
AN-33 UT	9	1036 ± 107	0.858 ± 0.057
AN-33 T + UT	21	1054 ± 56	0.850 ± 0.030
AN-42 T	12	1034 ± 30	0.800 ± 0.009
AN-42 UT	7	1038 ± 40	0.800 ± 0.013
AN-42 T + UT	19	1036 ± 23	0.800 ± 0.007
Ust'-Il'ya Formation			
401/6 T	6	1103 ± 99	0.840 ± 0.048
401/6 UT	6	1010 ± 150	0.857 ± 0.064
401/6 T + UT	12	1077 ± 88	0.839 ± 0.040
401/13 T	4	1014 ± 155	0.819 ± 0.053
401/13 UT	4	775 ± 184	0.869 ± 0.053
401/13 T + UT	8	1004 ± 175	0.811 ± 0.055
401/15 T	6	1051 ± 556	0.768 ± 0.110
401/15 UT	6	1100 ± 226	0.768 ± 0.053
401/15 T + UT	12	1113 ± 203	0.759 ± 0.043

(T) samples treated by 1N solution of ammonium acetate; (UT) untreated samples.

Thus, the size-variable clay fractions from shales (Gorokhov *et al.*, 1997; 2001) sampled from three stratigraphic levels of the Kotuikan section, i.e., from the Ust'-Il'ya Formation top and from basal parts of the upper and lower Yusmastakh subformations, reveal the same slopes of lines in plotted in the isochron diagrams without account for data points on leachates. Such a coincidence is not occasional. The alternative possibility that the coincidence is incidental, characterizing the isotopic mixing lines at three separate levels in the rock succession 600 m thick, which accumulated in different facies zones, seems unreal. We have to assume therefore that the line slopes reflect in fact the age of a certain epigenetic event that took place about 1000–1100 Ma ago and affected the whole sequence of pre-Vendian deposits in the Anabar massif. It is probable that this event represented repercussion of the Grenvillian orogeny manifested worldwide. The problem of duration or stages in the transformation history of rocks and of the sedimentation time proper remains open.

ON SOME ASPECTS OF METHODOLOGY

The paper we critically analyze here shows that its authors knew *a priori* the Yusmastakh Formation age

before the commencement of research, although they admitted that the "time range of the Yusmastakh Formation is inferable from a set of isotopic-geochronological, chemostratigraphic, and paleontological data, interpretation of which in current publications is not always unambiguous" (Gorokhov *et al.*, 2001, p. 6). Such an ambiguity can evidently be caused by different reasons, first, by the essence of used methods.

One of such modern methods used by the authors is the carbon isotopic geochemistry, which is of a wide application now in chemostratigraphic studies. Its usefulness for the Riphean rock successions is still doubtful in many aspects. First, the accepted reference curve of secular $\delta^{13}C$ variations is based on some isolated and incomplete reference sections, the stratigraphic subdivision of which is not always sufficiently substantiated (Raaben, 2001). Second, the Riphean time scale is very problematic, and each of three Riphean subdivisions exceeds in duration a half of the Phanerozoic. In the Phanerozoic portion of the curve (Veitzer *et al.*, 1999) compiled with a step of 20 Ma, the scatter of $\delta^{13}C$ values within every step exceeds that the Riphean isotopic chemostratigraphy is based on. Variations of $\delta^{13}C$ values in carbonate rocks of the Upper Proterozoic (500–800 Ma) are up to 8‰ within time intervals several tens of million years long (Jacobsen and Kaufman, 1999; Walter *et al.*, 2000). Therefore, it is hardly possible at the moment to lean upon C-isotope data when correlating the remote Middle and Lower Riphean sections, particularly taking into consideration the fact that variations in carbonate carbon isotope composition correspond to 1–2‰ in the entire pre-Vendian portion of the Kotuikan section.

Gorokhov and his co-authors paid also a keen attention to mineralogical study of clay fractions from examined samples, but some of their *a priori* postulates deserve critical comments, the main of which are as follows. (1) They believe that the 1M illites are always authigenic, but this is incorrect. The 1M modifications of micas are sufficiently stable and take part in sedimentation like other phyllosilicates. (2) When analyzing Riphean deposits, it is hardly possible to speak about authigenic minerals of the diagenesis zone. At any rate, it is necessary to prove that "structural perfection" of separated subfractions is not a result of sample treatment (trituration, ultrasonic machining, dissolving). (3) As is stated in the work, "differences in crystallinity indices of illite from fine and coarser subfractions may point to their origin at different stages of lithogenesis" (Gorokhov *et al.*, 2001, p. 10). However, this may be not the case. The difference between crystallinity indices of coarse- and fine-grained subfractions indicates first of all an easier abradability of a mineral phase with the larger half-widths of peak 001. As it was shown long ago (Kulikova *et al.*, 1966), the easier abradability of mineral phases leads to their concentration in most fine-grained fractions. (4) The discussed contribution of feldspars to the isotopic age of subfractions and the detrital origin of these minerals accepted *a priori* (Gorokhov *et al.*, 2001, p. 19) are incorrect at least. Phrases concerning authigenic or allogenic origin of feldspars from fractions <3 μm from Riphean rocks are simply senseless, because none of the origin can be proven.

As for difficulties in biostratigraphic subdivision and correlation of Riphean sections, they are well known (Semikhatov, 1995, 1997, and others), and the authors of the paper under consideration perfectly demonstrate stating that "the suggested Late Riphean age of organic-walled microfossils from the lower Billyakh deposits is unacceptable because of two reasons. First, it is incompatible with isotopic dates..." (Gorokhov *et al.*, 2001, p. 6). We break this sentence at the point very important in terms of methodology. It is evident that if the methods in use do not answer positively to the posed question, any quantity of the applied methods cannot change or clear the situation. One should lean upon the only method, which is accepted at present as decisive. As it follows from the cited sentence and general ideology of the paper, the authors consider the isotopic dating precisely as a method of this kind. However, interpretation of their own data is misleading from our standpoint probably because of their faith in correctness of inferences based on other investigation methods.

In fact, there is no contradiction between biostratigraphic, chemostratigraphic, and isotopic scales, because paleontologists view chronological relationships between sedimentary successions through the prism of organic world evolution only. Biostratigraphic units, which are defined using the most representative successions of microfossils, remain as such regardless of any isotopic dates assigned to them (e.g., Veis *et al.*, 2001).

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Once More about the Rb–Sr Isochron Method of Dating the Sedimentary Rocks

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INTRODUCTION

Over 40 years ago, Nicolaysen (1961) demonstrated that data points corresponding to several *cogenetic phases* of a system[1] plot in the $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagram along a straight line (an isochron), if the system was closed and evolution of $^{87}Sr/^{86}Sr$ ratios was controlled by ^{87}Rb radioactive decay only. Since that time, the isochron method was a routine of geochronology used to date both the mineral associations and the series of cogenetic whole-rock samples. A broad application of isochron method that is seemingly simple, when the computer-based assessment of isochrons is at hands, does not mean, however, that one can ignore or underestimate some basic principles of the method. Of prime significance among the latter are the conditions of approach to dating the rocks of various origin, which have been sampled in a different manner for the isotopic analysis (whole-rock samples, mineral parageneses, fine-grained fractions of various size ranges). The paper by Vinogradov *et al.* (2003) published in this issue of the journal is a proper occasion to discuss the problem once more.

The Rb–Sr isochron method was suggested to measure age of magmatic and metamorphic rocks based on data of isotopic analysis of their minerals or whole-rock samples. At present, the isotopic analysis of *minerals* from a single rock sample (plotting of a "mineral isochron") is a routine procedure of any isotopic method, the Rb–Sr method included, used to determine age of magmatic or metamorphic event under examination. As a rule, there is no doubt that minerals of a magmatic rock sample are cogenetic, although their crystallization times can be different to a minor extent. A more complex problem is to date particular metamorphic events by means of mineral isochrons, because we must use in this case the data of modern petrology to be granted that a mineral isochron under consideration characterizes a paragenesis of metamorphic minerals, which were in equilibrium at their formation time.

The isotopic dating of the whole-rock samples of magmatic and, especially, of metamorphic rocks is a more complicated task. We may suspect in this case, using the Rb–Sr method, that the calculated age can be either exaggerated or rejuvenated due to peculiarities of the pre-crystallization history of magmatic melts and protoliths, on the one hand, and because of a possible post-crystallization disturbance of isotopic systems in the samples under examination, on the other. Nevertheless, a comprehensive study of rocks by various methods, in addition to the isotopic ones, and clear understanding of specific features of rocks selected for isotopic dating open a possibility to avoid the undesirable results (see reviews in Gorokhov, 1985; Faure, 1986; and references therein).

As for sedimentary rocks, their specific composition means that the Rb–Sr isochron method can be applied to their minerals or whole-rock samples with a due precaution and proper insight into existing problems. Let us consider these problems bearing in mind that the *cogenetic status* of analyzed samples is a principal requirement of the isochron model. The other thing important therewith is a necessary to take into account the *independent* data of geology, mineralogy, and geochemistry when coming to interpretation of isotopic measurement results in terms of analytical geometry using the $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagrams.

Rb–Sr ISOCHRON METHOD AS APPLIED TO MINERALS OF SEDIMENTARY ROCKS

Most important sedimentary Rb–Sr geochronometers are authigenic minerals of siliciclastic rocks, such as globular glauconite, fine-grained illite, and mixed-layer clay minerals, which have high Rb/Sr ratios. Bonhomme and Millot (1966) from the Strasbourg University who were first to use the fine-grained (less than 2 μm) illite fractions in practice of isotopic dating dem-

[1] In terms of Rb–Sr systematics, these are phases with identical $^{87}Sr/^{86}Sr$ initial ratios.

onstrated a theoretical possibility of origination and transformation of clay minerals under conditions of sedimentation and diagenesis. In the last two decades, these and other experts in geochronology published many works on the isotopic systematics of clay minerals originated at various stages of lithogenesis, those of the Precambrian included (Bonhomme, 1982; Clauer, 1976, 1979, 1982a; Gorokhov and Semikhatov, 1984; Semikhatov and Gorokhov, 1984; Gorokhov, 1985; Clauer *et al.*, 1992; Clauer and Chaudhuri, 1995). As is established in particular, the deposition age of clay sediments "cannot be determined directly by the Rb–Sr method, because none of the mechanism at the stage of sedimentation is able to homogenize (provided the variability of Rb/Sr ratios) the Sr isotope composition in sediments as a whole or in any their fractions" (Gorokhov, 1985; p. 117). This conclusion is correct even in respect to abyssal red clay accumulated under extraordinary low sedimentation rate (Semikhatov and Gorokhov, 1984; and references therein). In distinction, the burial diagenesis responsible for transformation of mixed-layer clay minerals and for generation of voluminous elision fluids (Kholodov, 1983) is an event appropriate for the Rb–Sr dating based on the analysis of mineral fractions separated from clastic deposits. For the Precambrian, the burial diagenesis age can be regarded as a first approximation to the age of sedimentation.

In subsequent works aimed to perfect methods of dating the clay rocks and minerals, the attention was focused on separation and Rb–Sr analysis of fine-grained subfractions (Morton, 1985; Clauer *et al.*, 1990; Gorokhov *et al.*, 1994, 1997, 2001, 2002; Schaltegger *et al.*, 1994; Zwingmann *et al.*, 1999). Principles of this approach are identical to those of the "mineral isochron" method in application to crystalline rocks and require, first of all, separation of submicron clay fractions. The approach is aimed to concentrate, if not separate completely, the clastic and authigenic minerals of siliciclastic rocks in individual subfractions. Such a separation into subfractions is a first step toward dating the discrete stages of lithogenesis. In contrast to the statement by Vinogradov and his colleagues (2003), the Rb/Sr ratios in subfractions of an equal size range from a single sedimentary horizon may be different (Morton, 1985; Gorokhov *et al.*, 1994; Uysal *et al.*, 2001), and subfractions of this kind can be successfully used in some cases to date the stages of lithogenesis. Nevertheless, this is not a way to solve the problem completely, because the procedure does not yield, in the scale of a sample, the true *cogenetic* substances satisfying the isochron method requirements. These substances incompletely separable by mechanical or physical methods can be obtained by chemical leaching under appropriate conditions.

The practice of geochronology showed that leaching of fine-grained shale and clay fractions in different reagents is an effective procedure to separate soluble phases from insoluble layered silicates. The main difficulty here is identification of the material solved. We can apply various methods to analyze the mineral and chemical compositions of untreated fractions of layered silicates or of residues after leaching, but the only way to assess the composition and origin of the leached material is the chemical analysis.

It is obvious that the Sr isotope composition in a given leachate is integral, reflecting this parameter in soluble non-silicate mineral phases, on the one hand, and in exchangeable interlayer positions and basal surfaces specific of clay minerals. As is shown in a series of recent works (Toyoda and Masuda, 1991; Ohr *et al.*, 1991, 1994; Clauer *et al.*, 1993; Schaltegger *et al.*, 1994), a considerable part of leached material is deriver from associated soluble minerals (carbonates and phosphates), but not from the exchangeable positions of the clay minerals proper. This suggests that the Rb–Sr fractionation in sediments took place shortly after their deposition. As a result, Rb and Sr enter the authigenic phases, which are able to behave as closed geochemical systems and can be separated from each other by means of the laboratory leaching. If this is correct, the Rb–Sr dates calculated for pelites from the diagenesis zone based on the three-point "inner isochrons" (untreated subfraction–acid/acetate leachate–residue after leaching) can be regarded as equivalents of age values corresponding to mineral isochrons for authigenic carbonates (phosphates) and illite (mixed-layer illite-smectite). If the above opinion is incorrect, the obtained dates should be considered as model ages, which are calculated using the $^{87}Sr/^{86}Sr$ initial ratios assessed not quite correctly. As the calculated initial ratios are usually at covariance with the effective diameter of studied particles, the first alternative is preferable. Such a covariance indicates that subfractions of different size ranges are unequally enriched in illite formed under different conditions, which were dissimilar to those at the formation time of associated soluble authigenic minerals. By no means, it is an indication of a process that equilibrated the Sr isotope composition in associated phosphates and carbonates.

The three-point straight lines in the $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagrams have geochronological meaning under two necessary conditions: first, the leachate composition must correspond to that of the carbonate and/or phosphate phases, which exist in reality; and second, a selective extraction of radiogenic ^{87}Sr took no place in the course of laboratory leaching. The last problem appeared, when Faure and Barret (1973) established that $^{87}Sr/^{86}Sr$ ratios in muriatic leachates from nonmarine carbonate rocks are much higher than the corresponding ratios in seawater. Percentages of siliciclastic fraction in the rocks studied by these researchers were, however, extremely high (up to 83%) and correlative with the $^{87}Sr/^{86}Sr$ ratios in carbonate material. According to recent data (Semikhatov *et al.*, 2002), the high $^{87}Sr/^{86}Sr$ ratios in this material are likely a consequence of diagenetic exchange with the silicate admixture rather than the result of selective extraction of radio-

genic ^{87}Sr from coexisting clay minerals by leaching in 0.1N HCl, as Faure and Barret believed. Clauer (1979) also observed extraction of radiogenic ^{87}Sr from 1M illite treated in 2.5N HCl, but the later leaching of the same illite fraction in 1N HCl showed that the $^{87}Sr/^{86}Sr$ ratio in leachates is identical to that in the associated carbonates (Clauer *et al.*, 1993). As a result, the Sr isotope composition in 1N HCl leachates was rightfully regarded as characterizing the formation medium of clay and carbonate minerals.

There were other experiments with different leaching reagents, which could be able to extract selectively the radiogenic ^{87}Sr from clay minerals, if this would be possible. For instance, Clauer with colleagues used 1N HCl solution to leach the recent authigenic silicates of marine origin, such as phillipsite, nontronite, beidellite (Clauer, 1982b; Clauer *et al.*, 1982, 1984), and detrital smectite (Holtzapffel *et al.*, 1985; Clauer *et al.*, 1990). Kralik (1984) compared the Rb–Sr data for the fine-grained illite fractions leached in 1N HCl or in ammonium ethylendiaminetetraacete (NH_4–EDTA) and extracted by cation-exchange resin. Ohr *et al.* (1991) used 1N HCl and CH_3COOH to leach the clay fractions of Tertiary shales from a borehole drilled at the Mexico Bay coast. Clauer *et al.* (1993) studied the Rb–Sr systematics treating the fine-grained fractions of the Lower Cambrian blue clay from northern Estonia in a series of reagents, such as 1N HCl, 1N NH_4Cl, humic acid, acetone, and NH_4–EDTA, and filtering solution through a cation-exchange resin. None of these works demonstrated an effect of selective leaching of radiogenic ^{87}Sr from clay fractions bearing the 1M illite. Therefore, Vinogradov and his colleagues are incorrect from the geochemical standpoint, when they state that the basic mass of radiogenic Sr is selectively extracted from clay fractions of the Yusmastakh shales treated in ammonium acetate solution with pH = 7. Another assumption that Rb and Sr enter the leachates as a result of congruent dissolution of clay minerals is incompatible with slopes of lines drawn in the $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagrams through data points, which characterize the acetate leachates from subfractions of different size ranges (Gorokhov *et al.*, 2001, Fig. 9). The fact is that the Riphean succession of the Anabar massif, in which the Yusmastakh Formation is terminal one, is separated by a principal unconformity from pre-Riphean magmatic complexes of the crystalline basement, the geologically youngest rocks of which yielded the U–Pb zircon ages of 1843 ± 12 and 1760 ± 20 Ma (Rozen *et al.*, 2000; and references therein). Dates equal to the inclination of lines in Fig 9 (Gorokhov *et al.*, 2001) our opponents refer to correspond to 1950–2300 Ma and greatly exceed the maximum age limit of the Yusmastakh Formation.

Thus, the leachochrons for fine-grained subfractions of clayey rocks are the analogues of mineral isochrons for magmatic and metamorphic rocks but not "the artificial means of age calculation," as they are termed by Vinogradov and his colleagues. The weakness of arguments in favor of the last viewpoint is likely clear for our opponents themselves, because they write more carefully, in distinction from their common categorical statements, that "the Sr easily soluble from a sample *may have and sometimes actually has nothing to do* (italicization is ours) with the Sr hard-bonded in the crystal lattice of minerals" (Vinogradov *et al.*, 2003, p. 638). Moreover, in their work devoted to isotopic systematics of pre-Vendian deposits of the Baikit Uplift, our opponents used themselves the procedure of leaching shales in HCl and pointed out (Vinogradov *et al.*, 1998a) that "this treatment may be useful by assessment of the doubtful errorchrons' significance." We can add to the aforesaid that the real geochronological meaning of leachochrons is evident from their concordance with Pb–Pb isochron dates obtained for associated carbonate rocks (Ovchinnikova *et al.*, 1998) and/or with conclusions based on data of paleontology and C-isotope chemostratigraphy (Knoll *et al.*, 1995; Gorokhov *et al.*, 1997, 2001; Semikhatov *et al.*, 2002).

Let us turn now to considerations of Vinogradov *et al.*, which evidence, as they say, the "doubtless" Sr isotopic equilibration in the whole-rock samples of non-metamorphosed siliciclastic sediments or between their fine-grained subfractions of clay minerals.

Rb–Sr ISOCHRON METHOD AS APPLIED TO FINE-GRAINED SUBFRACTIONS OF VARIOUS SIZE RANGES FROM SILICICLASTIC ROCKS

In Figs. 1 and 2 of their paper, Vinogradov and his colleagues (2003) reproduce our Rb–Sr data for clay subfractions of various size ranges from the Ust-Il'ya and Yusmastakh formations of the Anabar massif (Gorokhov *et al.*, 1997, 2001), stating therewith that the established linear arrangement of data points is of the isochron type. Their statement is unacceptable even formally, because it contradicts the main principles of the isochron method. In fact, the proving system presented in our publications clearly shows that both a clay material of fine-grained subfractions and the mobile phase extracted from them after leaching in the ammonium acetate solution consist of mixed non-cogenetic minerals (Gorokhov *et al.*, 1997, Fig. 6; Gorokhov *et al.*, 2001, Figs. 9 and 10). Discarding this conclusion, Vinogradov *et al.* (2003) suggest that the mobile Sr extracted by ammonium acetate from clay subfractions has no connection with the Sr hard-bonded in the crystal lattice of minerals. Whoever may be right, it is impermissible to use our data for untreated subfractions in the isochron plots because of two reasons. First, each of the subfractions consists *a priory* of non-cogenetic phases bearing both the mobile and hard-bonded Sr. Their sentence that "a minor percentage of substance entering the leachate does not disturb the substance balance in a given sample" (Vinogradov *et al.*, 2003, p. 639) is a disinformation, since one can easily see from our Table 2 (Gorokhov *et al.*, 2001) that "a minor percent-

age" can be as great as 22–33% (!) of the total Sr content in a subfraction. Second, it is incorrect to unite data points, which characterize the non-cogenetic untreated subfractions and corresponding residues after leaching, as it is done in Figs. 1 and 2 of the paper under discussion. In this case, our opponents ignore basic principles of the isochron method, simultaneously using it to prove their own viewpoint.

To substantiate our conclusion that clay subfractions from shales of the Yusmastakh Formation consist of two phases at least and to show variable proportions between these phases in different subfractions, we considered in our work (Gorokhov *et al.*, 2001) a series of chemical and mineralogical (X-ray diffraction data) arguments. Our opponents view these arguments as unconvincing, thus being in a flagrant contradiction with the worldwide practice in the sphere of clay mineralogy. They also tend to convince a reader that different crystallinity indices of studied illites are the result of sample trituration in laboratory rather than the indication of their formation at different stages of lithogenesis. These researchers likely forgot their own work (Vinogradov *et al.*, 1999; p. 401), in which they used the same criterion to assess the alteration degree of rocks they studied. In addition, we should stress that our method of separating clay fractions does not include at all the trituration stage, as explained in both works devoted to geochronology of Riphean shales from the Anabar massif (Gorokhov *et al.*, 1997, 2001). Therefore, this factor must be simply excluded from consideration, when our data are under examination. To support once more the two-phase composition of clayey material from shales of the Yusmastakh Formation, we may add here to the former arguments a new one: K/Rb ratios are essentially different in coarse- and fine-grained illite subfractions we studied (380–630 versus 160–200). According to Clauer *et al.* (1993), this fact alone indicates the difference in clay mineralogy of subfractions having different size ranges.

In a series of their recent works, Vinogradov *et al.*, (1994, 1998a, 2000a, 2000b) paid a serious attention to mixing lines and considered some examples of such pseudo-isochrons established for Pliocene–Quaternary mud, which is composed of mutually precipitated volcanogenic and terrigenous materials (Asahara *et al.*, 1999), and for different-size fractions of recent marine siliciclastic sediments (Eisenhauer *et al.*, 1999). Therefore they should clearly understand that any Rb–Sr diagram plotted for sediments of variable grain sizes depicts the mixing lines lacking any geochronological sense. Nevertheless, our opponents interpreted the linear trends, which characterize different-size fine-grained subfractions from Riphean shales of the Ust-Il'ya and Yusmastakh formations, as errorchrons (Vinogradov *et al.*, 2003, Figs 1 and 2) reflecting "a certain epigenetic event that took place about 1000–1100 Ma ago and affected the whole sequence of pre-Vendian deposits in the Anabar massif." It is not important for them that different mineral and chemical compositions of these fine-grained subfractions seriously limits a possibility of Sr isotopic equilibration between the subfractions. It is clear that diffusion in solid phases is ineffective under low temperature, and the epigenetic isotopic equilibration is attainable therefore after a complete recrystallization of clay minerals under influence of interstitial fluids. Such a recrystallization that is necessary in opinion of Vinogradov *et al.* (2000b) must lead not only to the Sr isotope homogenization, but also to equalization of mineral and chemical compositions of clayey material in subfractions that has not been detected in the studied shales. We should repeat here once more (*repetitio est mater studiorum*) that negative correlation between Rb/Sr ratios and clay particle sizes, which is established for shale samples from the Yusmastakh Formation (Gorokhov *et al.*, 2001), excludes a possibility of illite formation in the course of bulk crystallization or recrystallization at any stage of lithogenesis. Consequently, the idea of Sr isotopic equilibration in clay subfractions of shales, which is suggested by Vinogradov *et al.* and seems attractive from the standpoint of analytical geometry, is unable to stand an elementary geochemical test.

Let us explain now why linear trends in Figs. 1 and 2 (Vinogradov *et al.*, 2003), which are not the isochrons, have similar slopes. As one can easily see from the Figures, the highest $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ ratios characterize those data points, which correspond to residues after leaching of finest (<0.1 and 0.1–0.2 μm) clay subfractions (Gorokhov *et al.*, 2001, Figs. 5 and 6). Exactly these subfractions are most rich in younger illite generations 1000–1200 Ma old. In accord with the least-squares method, slopes of linear trends highly depend on positions of extreme data points, and it is not surprising therefore that dates calculated for the aforementioned pseudo-isochrons are close to the formation time of younger illite generations. Surprising is another moment. Considering linear trends in the $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagrams, Vinogradov and his colleagues point out that the trends slopes "are specified by data points far remote from the origin of coordinates," and that "the calculated age can be close to a certain reality." Thus, they apply a double standard to interpretation of the Rb–Sr diagrams. Understanding an artificial character of age values presented in the table, our opponents state nevertheless that positions of lines in Figs. 1 and 2 "leave no place for a randomness," and their slopes "really depict the age of some epigenetic event." In fact, the $^{87}Sr/^{86}Sr$ ratios measured for residues after leaching are so high that the apparent age values are almost insensitive to $^{87}Sr/^{86}Sr$ initial ratios, which may be accepted. In this case, therefore, we deal with the model but not isochron ages of younger (1000–1200 Ma old) illite generations. In distinction from our opponents who attribute the latter to "some epigenetic event," we discussed in our works (Gorokhov *et al.*, 1997, 2001) the concrete geological processes, which might be and were responsible for origin of these illites. Thus, it is clear that the pseudo-isochrons of our oppo-

nents are incorrect in terms of the isotopic geochronology and useless with respect to problems of lithogenesis in Riphean deposits of the Anabar massif.

Rb–Sr ISOCHRON METHOD AS APPLIED TO WHOLE-ROCK SAMPLES OF SILICICLASTIC DEPOSITS

We would like to mention first that Vinogradov and his workmates are nowadays the only researchers who apply the Rb–Sr dating method to *whole-rock samples* of fine-grained siliciclastic deposits, although the latter have been originally regarded as promising geochronometers (Compston and Pidgeon, 1962; Whitney and Hurley, 1964). The priority of clayey rocks was evident from high K and Rb concentrations in their main constituents, the layered silicates, and it was originally *suggested* that isotopic exchange with seawater leads to equilibration of the Sr isotopic composition in the rocks, making them appropriate for dating by the Rb–Sr isochron method. The situation changed after classical works by Dash and his colleagues (Dash *et al.*, 1966; Dash, 1969) who showed that neither such an exchange, nor the intermixing of heterogeneous clastic material are able to homogenize the Sr isotope composition in marine sediments, concrete samples of which retain different Rb/Sr ratios. In subsequent works devoted to dating of siliciclastic rocks, the attention was paid, therefore, not to the whole-rock samples (see review in Semikhatov and Gorokhov, 1984; Gorokhov and Semikhatov, 1984). Only the group headed by Cordani was an exception for a while (Cordani *et al.*, 1978; Misuzaki *et al.*, 1994).

As is known, clastic and authigenic components of terrigenous rocks interact differently with environmental medium in the course of postsedimentary processes, and corresponding whole-rock samples are therefore inappropriate for dating by the Rb–Sr isochron method. Since relevant problems have been already considered in detail (Clauer, 1982a; Gorokhov and Semikhatov, 1984; Gorokhov, 1985; Clauer and Chaudhuri, 1995), we discuss below only some necessary aspects.

Siliciclastic deposits are the mixtures of minerals different in age and genesis and formed at different stages of lithogenesis. Data points characterizing their whole-rock samples must commonly plot in the $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagrams outside the "reference isochrons," corresponding in age to a postsedimentary process that did not result in a complete reworking of sedimentary material. Positions of these data points relative to the reference isochrons is defined by the following relation:

$$\frac{\text{original clastic material} + \text{partially reworked clastic material}}{\text{completely reworked clastic material} + \text{newly formed material}}.$$

Perry and Turekian (1974) who studied a thick Miocene succession of clayey sediments of the Mexico Bay coast showed long ago that the burial diagenesis is unable to homogenize, even at the depth of 5 km, the Sr isotope composition in the rock fractions containing detrital minerals (feldspar, illite, kaolinite) in different proportions. Nobody can expect therefore such a homogenization in sedimentary bodies composed of shales, siltstones, and sandstones, which are up to dozens and hundreds meters thick and extending for dozens and hundred kilometers in the lateral direction. Meanwhile, Vinogradov *et al.* (1994, 1996, 1998a, 1998b, 1999, 2000a, 2000b) sampled rock bodies of this kind exactly, attempting to determine the Rb–Sr age of epigenetic processes in Riphean successions of the Urals and Siberia. A good example is their analysis of Rb–Sr data obtained for shales of the Ust-Il'ya Formation of the Anabar massif, which are illustrated in Fig. 2 and presented in the Table of their work (Vinogradov *et al.*, 2003). As they believe, linear trends in Fig. 2 are errorchrons but not the mixing lines and depict the age of "some epigenetic event that ... affected the whole pre-Vendian succession of the Anabar massif." In such a case, we can expect the homogenized Sr isotope composition in the fine-grained subfractions of shale samples collected a few meters apart from each other in a single bed. Data presented in the table show, however, that "initial" (as they are termed by our opponents) $^{87}Sr/^{86}Sr$ ratios in clay fractions of various samples are noticeably different. Consequently, it is impossible to speak here of the Sr isotope homogenization neither in the regional sense ("the whole pre-Vendian succession of the Anabar massif"), nor with respect to whole-rock samples collected from particular formations. Thus, the postsedimentary transformations in siliciclastic rocks, which are unable to alter completely clastic feldspars and micas, are similarly unable to homogenize the Sr isotope composition in all phases of the rock and *a fortiori* in the whole-rock samples of any volume, which are collected far apart. This conclusion was done over two decades ago (e.g., Clauer, 1976) and forced researchers to find new ways in seeking for geochronological information about processes of lithogenesis.

The way chosen by Vinogradov *et al.* (1994, 1996, 1998a, 1998b, 1999, 2000a, 2000b) is the Rb–Sr analysis of whole-rock samples of siliciclastic deposits with subsequent interpretation of analytical results using the $^{87}Rb/^{86}Sr$–$^{87}Sr/^{86}Sr$ diagrams. Having chosen the isochron method as a basic, if not a single tool of investigations, these researchers do not meet the main conditions of its application in many of their works. Regardless of the worldwide practice, they believe without substantiation that linear trends, plotted in those diagrams for non-cogenetic samples and assessed with a considerable dispersion (the so-called errorchrons), are appropriate for age determination of geological processes. A term used to designate these processes is "epigenesis," i.e., the term of a broad meaning, not implying an asso-

ciation with concrete geological events. In some cases, however, when clastic material is intensely reworked, dates calculated for whole-rock sedimentary samples may approximate to the ages of real processes.

The experience of Rb–Sr dating of siliciclastic rocks different in age shows that physical and mineral transformations in these rocks begin at the stage of anchimetamorphism (metagenesis) and last until the formation stage of stable mineral parageneses of the greenschist or amphibolite metamorphic grade (Peterman, 1966; Montigny and Faure, 1969; Gorokhov *et al.*, 1982; Graham, 1985; Gorokhov, 1987). Under such conditions, metapelites, even if they are sampled throughout the distribution region (Gorokhov, 1985) with sampling sites set as far apart as dozens of kilometers, may yield sometimes the Rb–Sr isochrons bearing geological meaning. This does not imply of course that substance really migrated for such a distance. The whole volume of a protolith can be divided into a series of partial, regularly distributed volumes. If the latter have identical Rb/Sr ratios equal to the average one in the protolith, then the average $^{87}Sr/^{86}Sr$ ratios in each partial volume will be also identical, provided that the ^{87}Rb radioactive decay is the only controlling factor (Roddick and Compston, 1977; Gorokhov, 1985).

Taking the aforesaid into account, we admit that the Rb–Sr errorchron dates (880–980 Ma) calculated by Vinogradov *et al.* (1999) for the whole-rock shale samples from the Lower Riphean Burzyan Group (1650 ± 50 to 1350 ± 50 Ma old) of southern Urals are quite rightful. They may reflect *a tendency* to the Sr isotopic equilibration in the course of epigenetic alterations in the Burzyan succession, especially because the middle part of the latter hosts a large siderite deposit, the metasomatic origin of which 1010 ± 100 Ma ago was accompanied by substance mobilization in the host siliciclastic rocks (Kuznetsov *et al.*, 2001). It is necessary to stress, however, that a complete homogenization of the Sr isotope composition in whole-rock samples from the Burzyan Group, which have been analyzed by Vinogradov *et al.* (1999), is hardly admissible. First, all the samples include clastic feldspars, micas, and fragments of allogenic rocks in addition to a variable amount of authigenic minerals enriched in K and Rb. Second, the crystallinity index of illite implies that postsedimentary alterations in the rocks took place within the zone of anchimetamorphism (Vinogradov *et al.*, 1999, pp. 402 and 409). It is incorrect to state therefore that rocks of the Burzyan Group are locally metamorphosed under the high-temperature conditions (550°C) of the amphibolite facies (Vinogradov *et al.*, 1999, p. 410) or biotite (sub)facies (Vinogradov *et al.*, 2000b), because the statement is inconsistent with real mineral parageneses (Maslov *et al.*, 2001). Moreover, when percentages of unaltered clastic material are high in the rock samples selected for the age determination, the errorchron approach is incorrect, internally contradictive, and inappropriate for geochronological practice. Let us consider several examples.

In their works devoted to isotopic analysis of Riphean deposits in the Uchur-Maya region of East Siberia, our opponents rightfully stated that "K–Ar dating of whole-rock samples of terrigenous sediments may lead to diverse results" (Vinogradov *et al.*, 1998b, p. 640), which may have "no connection with geologically significant events" (Vinogradov *et al.*, 2000a, p. 176). They forget to add therewith that this disadvantage is characteristic of the Rb–Sr method as well, because mineral geochronometers are the same in both cases. However, graphic interpretations used in the Rb–Sr systematics may create, without critical assessment, an illusion of reliable results, which open space for speculations.

Many contradictions in the works under discussion appear because of geochronological accents attributed to the Rb–Sr errorchron dates without proper substantiation. For instance, it was stated in one of the works: "the K–Ar system in clay minerals is more stable than the Rb–Sr system" (Vinogradov *et al.*, 1998b, p. 641), and a few lines below one can read that "more frequently these two systems yield the concordant ages." As is noted in the same paper slightly before, the K–Ar dates for glauconites and whole rock siltstone samples are close (though exceeding the Rb–Sr "ages" calculated by the authors), and this coincidence is declared, without any substantiation, as a result of "rejuvenation." In addition, authors of the work conclude, disregard of the worldwide practice: "age determined for glauconite cannot characterize the formation time of sediment or the time of the early diagenesis" (Vinogradov *et al.*, 1998b, p. 640). Trying to explain in their other work why the K–Ar age of glauconite turned out to be older than the "age" of whole-rock shale samples, Vinogradov *et al.* (2000a, p. 179) formulated an incredible assumption that glauconite lost K but retained radiogenic Ar.

As was established, the Rb–Sr errorchron dates for silty and sandy shales from four Riphean Formations of the Uchur-Maya plate are increasing downward the succession: 690 Ma in the Upper Riphean Kandyk Formation, 811 and 810 Ma in the Middle Riphean Omnya and Konder formations, and 954 Ma in the Lower Riphean Omakhta Formation. In opinion of Vinogradov *et al.* (2000a), the dates indicate that clay minerals of sediments constituting the formations experienced an intense transformation, when they crossed some critical levels of the depth and temperature in the course of burial. Analyzing data on possible *P–T* conditions of such a transformation that could be regarded as explanation for the aforementioned Rb–Sr errorchron dates, they concluded that it took place under the temperature of 80–100°C at the depth of 2.5–3 km, although parameters assessed by other researchers are higher (260 ± 30°C at the depth of about 5 km; Vinogradov *et al.*, 2000a; and references therein). This interpretation of the Rb–Sr errorchron dates is inconsistent with the following three issues.

First, anyone who knows Riphean geology of the Uchur-Maya region cannot accept the idea that southern margin of the Uchur depression (sampling area of the Omakhta Formation) and northwestern flank of the Omnya Uplift (sampling area of the Omnya and Konder formations) subsided 950–800 Ma ago down to the indicated depth. This is well evident from the thickness of overlying Riphean deposits (Semikhatov and Serebryakov, 1983; Shenfil', 1991). Second, color indices (Hayes *et al.*, 1983) of Riphean organic-walled microfossils known from the Uchur-Maya plate (Veis *et al.*, 1998; Veis and Vorob'eva, 2002) suggest that their host deposits have been never heated above 60–80°C. Third, the Yudoma-Maya depression that is adjacent to the Uchur-Maya plate on the east subsided in the Riphean deeper than the latter, but the Rb–Sr errorchron ages calculated by Vinogradov *et al.* (2000a) for siliciclastic deposits of that depression are independent of their stratigraphic position and change irregularly, although they should steadily increase downward exactly here, as one could expect from the model under consideration.

Works by Vinogradov *et al.* (1994, 1998), which are devoted to the Rb–Sr systematics of Riphean deposits in inner areas of the Siberian platform (Katanga Depression and Baikit Uplift), are based on the analysis of occasional whole-rock samples of sandy and silty shales from various stratigraphic levels of boreholes. The latter are drilled as far apart as a few dozens or hundreds of kilometers, and it is quite unlikely that any geological process could be able to cause the Sr isotopic homogenization in such an impressive volume of non-metamorphosed siliciclastic deposits. Moreover, the analyzed samples of deposits bear clasts of feldspars, micas, and allogenic rocks besides the globules of authigenic glauconite that occurs sporadically and retains individuality of its Rb–Sr system. Disregard of all these facts, Vinogradov *et al.* (1994) attributed a geological meaning to the linear trend with an apparent age of about 1530 Ma, which has been plotted in the Rb–Sr diagram for deposits of the Katanga Depression. Investigating rocks of the Baikit Uplift (Vinogradov *et al.*, 1998a), they united in the diagram data points for 19 shale samples from various stratigraphic levels, which plot as an extended "cloud" (that is quite natural), and interpreted this trend as an evidence in favor of transformation of the whole pre-Vendian sedimentary cover of the region 1200–1300 Ma ago, thus suggesting an older age of the cover itself. None of their conclusions is consistent with data of paleontology (microfossils and stromatolites) and C-isotope chemostratigraphy, which clearly indicate that terminal horizons of pre-Vendian successions in inner areas of the Siberian platform are represented by Upper Riphean deposits (Khomentovskii and Nogovitsyn, 1998; Khabarov *et al.*, 2000) less than 1030 Ma old (Semikhatov *et al.*, 2000).

SOME PROBLEMS OF METHODOLOGY

In conclusion, we would like to discuss some problems of methodology used by Vinogradov and his colleagues, when they interpret their own results and criticize approach or conclusions of other researchers they disagree with.

The belief of our opponents that the main or single criterion, which validates the isotopic age determinations, is the linear arrangement of data points in the Rb–Sr evolution diagrams, leads to many contradictions in their works. However, all experts in isotopic geochronology know well that neither Rb–Sr nor K–Ar formal systematics are unable to prove or disprove in themselves the geochronological authenticity of the information obtained (the only exception is diagnosis of mixing lines). As for the mathematical apparatus used by Vinogradov *et al.* (1994, 1998a, 1998b, 1999, 2000b) in their geochronological works, we should mention, in particular, that Sokolov and Bujakaite (1986) clearly defined the application sphere of the mathematical statistics to analytical data: "A high authority of mathematics leads to some overestimation of the statistic method significance. In reality, the apparatus of mathematical statistics can play a very important, but certainly an *auxiliary* role and *cannot* replace (italicization is ours) a comprehensive geochronological and geological consideration of a given situation." This definition emphasizes that the linear arrangement of data points in the isochron diagram is a necessary but not the *sufficient* evidence in favor of geochronological and, especially, of geological significance of the calculated age value.

We should mention as well that it is always necessary to indicate by isotopic-geochronological considerations an extent of deviation of the analytical data obtained from the used mathematical (isochron) model. Our opponents follow this rule not everywhere however (e.g., Vinogradov *et al.*, 1996, 2000a, 2003). As compared to the isochron, the errorchron uncertainty is greater than experimental errors, thus being indicative of the isochron model inapplicability for the mathematical processing of analytical data. Therefore, the age values calculated based on the errorchron slopes bear a meaning only if there are *independent* arguments in favor of corresponding geological events. If the statistical uncertainty of an errorchron is missing, the latter lacks a geochronological meaning.

The mean square of weighted deviates (MSWD) is a parameter calculated by the polynomial least-squares method used to assess the data points scattering relative the approximation line. All experts in geochronology calculate this parameter using the well-known and clearly described procedure: evaluation of statistical weights of particular data points and deviation of measured values from calculated ones in the isochron diagram (McIntyre *et al.*, 1966; Williamson, 1968; Shukolyukov *et al.*, 1974). Unfortunately, parameters calculated in accord with this procedure cannot be com-

pared directly with the results obtained by our opponents, because they, trying to prove their geochronological models, calculate MSWD using the doubled standard deviations, illegally retaining the above abbreviation and terming the sought value as the root-mean-square deviation (Vinogradov *et al.*, 1998a, p. 272). One can easily be certain that such a substitution quarters the MSWD values and displaces the conventional boundary between isochrons and errorchrons toward the level convenient for our opponents.

As for a possible geochronological justifiability of the isotopic information in itself, we should mention that one of a few criteria enabling discrimination between mixing lines and age-meaning trends in the Rb–Sr diagrams is the test of 1/Sr correlation versus $^{87}Sr/^{86}Sr$ (Faure, 1986). Our opponents bring a reader to a nonplus however, when they consider their results obtained for the Precambrian deposits of the Yudoma-Maya depression and state that such a correlation evidences a geochronological meaning of calculated dates but not the two-phase mixing in the analyzed material (Vinogradov *et al.*, 2000a, p. 178).

It is also remarkable that Vinogradov and his colleagues reject a considerable part of their own analytical data in order to get the necessary linear trends in the Rb–Sr diagrams. For instance, they condemned, without specification of mineralogical or other arguments, 9 of 21 and 14 of 34 analyzed samples when calculated respectively the Upper Precambrian age of the Ura Uplift deposits (Vinogradov *et al.*, 1996) and the Lower Riphean age of deposits sampled in the southern Urals (Vinogradov *et al.*, 1999). In these cases, they acted indeed in accord with the ironic consequence from the Mayerthy law that reads as follows: any experiment can be regarded as successful, if less than 50% of measurements should be discarded to attain the consistence with a theory.

Turning now to the more general problems, we would like to stress that any isotopic-geochronological study must include a series of necessary interconnected stages. It is necessary (1) to define a problem; (2) to select samples enabling its solution; (3) to establish a degree of cogenetic origin of geochronometers based on the comprehensive compositional analysis of selected samples; (4) to get geochronometers rid of heterogeneous phases as completely as possible; (5) to use for isotopic analysis only phases enriched in cogenetic material; and (6) to consider all data of geology, mineralogy, chemostratigraphy, and paleontology, which may elucidate the real significance of calculated isotopic ages. In our opinion, the point most important here is information on the genesis and transformation conditions of deposits under dating.

To our regret, this is not the way Vinogradov and his colleagues follow completely in their isotopic-geochronological works. First, the deposits they analyzed to determine the isotopic ages are sampled at random, as is repeatedly emphasized in the works. Accordingly, the dated rocks are discretely distributed in the successions and over the study areas, and the collections studied include an uncontrollable proportion of samples inappropriate for solution of the problem posed (e.g., the non-metamorphosed arkosic shales used to date the low-temperature postsedimentary processes). Second, their works are lacking the necessary compositional characterization of dated samples and attempts to clean the latter from non-cogenetic mineral phases. Describing the dated samples, they tend mainly to prove that the chosen rocks are rich in K- and Rb-bearing minerals formed at the late diagenetic stages and leave aside the original mineral peculiarities of siliciclastic fraction. Third, Vinogradov and his colleagues do not consider properly the geological, tectonic, and mineralogical data elucidating (or maybe elucidating) the formation and transformation history of deposits, the whole-rock samples of which are used as geochronometers. Instead, they assess a geological significance of isotopic dates based on the statistic parameters of linear trends approximating data points in the isochron diagrams. Ending the discussion of the research methodology and polemics initiated by our opponents, let us touch shortly upon some other moments.

Terminology they apply describing the deposits under investigation implies operations in the sphere of chronostratigraphy. From this side of view, their statement that isotopic-geochronological data represent the main criterion used to define the Middle–Upper Riphean boundary in the Uchur-Maya region (Vinogradov *et al.*, 1998b, p. 633; Vinogradov *et al.*, 2000a, p. 179) is unexpected. Such a sudden jump into the chronometric sphere cannot be explained by an occasional slip of pen, because the sentence is repeated twice without any explanation of this jump and subsequent return in the chronostratigraphic sphere (Vinogradov *et al.*, 2003, p. 641). How deliberately they skip from one to another concept of the Proterozoic subdivision, which have nothing in common (Semikhatov, 1995), is the question to be asked in this connection.

Criticizing us for the use of data of C-isotope chemostratigraphy for substantiation of the Yusmastakh Formation age, Vinogradov *et al.* (2003) correctly remark that these data are useless for *correlation*, because the amplitude of $\delta^{13}C$ variations in the Riphean succession of the Anabar massif is within the limits of $0 \pm 1‰$ only (Pokrovskii and Vinogradov, 1991; Knoll *et al.*, 1995). However, we (Gorokhov *et al.*, 2001) used those data to assess the minimum age limit of deposits in question, but not to correlate the latter, as it is well known that the Riphean strata older than 1.25–1.27 Ga show the almost invariable $\delta^{13}C$ values close to zero, whereas this parameter is much more variable in overlying deposits (Knoll *et al.*, 1995; Kah *et al.*, 1999; Bartley *et al.*, 2001).

Vinogradov *et al.* (2003) also ascribe to us an opinion that we consider the isotopic-geochronological method as the principal one for age determination of the

Riphean microbiotas and, "proving" this, they cite a passage from our work (Gorokhov *et al.*, 2001, pp. 6–7) stopping in the middle of a sentence. The phrase fragment cited apart from the paper context may create an impression that our opponents are right, although it is clear from the whole sentence and from other our publications that we determine ages of concrete microbiotas using a complex of chronostratigraphic data, the vertical succession of characteristic taxa in the type and reference sections of the Riphean included. Exactly this approach is used to define the age position of microbiota that is under consideration in pages 6 and 7 of our paper.

Finally, it is stated in their work (Vinogradov *et al.*, 2003, pp. 640–641) that we (Gorokhov *et al.*, 2001) "knew *a priori* the Yusmastakh Formation age before the commencement of our research." Such a statement may be taken at the first sight for a reproach in disguise that we falsified our analytical data in order to get a consistence between the preformed opinion of the formation age and the Rb–Sr dating results. To say frankly, we are far apart from such a thought and believe that our critical comments will help Vinogradov and his colleagues to advance toward the standard approach to investigation of the K–Ar and Rb–Sr systematics of siliciclastic rocks, which is applicable in geochronological practice worldwide, in this country and abroad.

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In Searching for Global Stratotype Sections and Points of the Triassic Stage Boundaries

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The Subcommission on Triassic Stratigraphy (STS) of the International Stratigraphic Commission (ISC) approved the scheme of the Triassic subdivision into seven stages (Induan, Olenekian, Anisian, Ladinian, Carnian, Norian, and Rhaetian) by overwhelming majority at the Symposium in Lausanne, October, 1991, and endorsed it at the meeting in Kyoto, August 31, 1992 (Visscher, 1992). Almost all the listed stages have, however, semiofficial names, since they, except for the Induan, have no Global Stratotype Sections and Points (GSSP) officially approved for their lower boundaries. The Subcommission on Triassic Stratigraphy and its working groups are engaged at present in solving this problem.

On the suggestion of the STS, the Executive Committee of the International Union of Geological Sciences (IUGS) approved the first and yet the only of these boundaries–the lower boundary of the Triassic–in March, 2001 (Orchard, 2001). It is defined now at the base of Bed 27c of the Meishan Section D in the Changxing County (Zhejiang Province, southern China), where the first appearance of conodonts *Hindeodus parvus* Kozur et Pjatakova is recorded (Yin *et al.*, 2001), i.e. above the *Otoceras* Beds (Baud, 2002), which were traditionally taken for the Triassic base in the biochronological ammonite standard (Shevyrev, 1999, 2000, 2001). Being a subject of pride for Chinese geologists and a pilgrimage site for curious tourists, this section is under the state protection. A memorial plaque is mounted near it. Having the position of the global Triassic basal boundary approved in the Meishan section, the IUGS thus defined officially the base of the Induan as well. Proposed for candidates of the GSSP of the Olenekian lower boundary are sections in the southern Primor'e (Zakharov, 1994, 1994, 1996; Zakharov *et al.*, 2002), eastern Verkhoyansk region (Dagys, 1995, 1997; Shevyrev, 2002), Taimyr Peninsula (Dagys and Sobolev, 2000), and in the Anhui Province of southern China (Tong *et al.*, 2002; Zhao *et al.*, 2002).

Substantiating their choice, adherents of the first alternative argue that the ammonoid assemblages from Induan–Olenekian boundary deposits of the southern Primor'e include Boreal taxa making easier the interregional correlation of these deposits. Their opponents, upholding the second and third alternatives, point to the fact that the Olenekian Stage was first established in Siberia, where precisely the GSSP for that boundary should be selected. Stratigraphers from China believe that the GSSP for the Olenekian lower boundary should be sought for in the Tethyan paleobiogeographic realm, where all other Triassic stages have their stratotypes. In their opinion, the western Pingdingshan section in the Anhui Province, where this boundary is defined at the first appearance level of conodonts *Neosphathodus waageni* Zweet above the *Flemingites* Zone base of the ammonite standard is a best candidate for the GSSP in question (Tong *et al.*, 2002).

Since the Induan–Olenekian boundary has not yet been approved officially and corresponding stage ranges still remain disputable, Chinese stratigraphers suggest to use instead the new stages: the Inkenian Stage with the stratotype in the Meishan section of the Zhejiang Province and the Chaohuan Stage with the stratotype in the Majiashan section of the Anhui Province (Tong *et al.*, 2001). The western Pingdingshan section mentioned above is considered therewith as suitable for the boundary stratotype.

Thus, there are two pairs of alternative stage names (Induan versus Inkenian and Olenekian versus Chaohuan), and at least four GSSP candidates suggested for their boundaries (in the southern Primor'e region, the eastern Verkhoyansk region, in Taimyr Peninsula, and in southern China). Thus, the International Working Group on the Induan–Olenekian Boundary has to make a difficult choice.

The International Workshop held by the Romanian Academy of Sciences at the Bucharest University, and the STS meeting in Tulcea (northern Dobrogea), June 7–10, 2000, was devoted to the problem of the Lower–Middle Triassic boundary, in other words, of the boundary between the Olenekian and Anisian stages (*Workshop...*, 2000). The main objective of the field trip held in the first day of the workshop was to observe two partially overlapping sections of upper Olenekian and lower Anisian deposits in the southern slope of the Desli Caira Hill situated south of Tulcea. One of the

sections exposed in the slope eastern part is suggested by Romanian geologists as a possible candidate for the GSSP of the Anisian lower boundary (Gradinaru, 2000). It is composed of massive thick-bedded Hallstatt-type limestones reddish in color and more than 60 m thick. Eleven horizons with ammonoids were distinguished in the upper half of this sequence (38 m). The Olenekian–Anisian boundary is defined between the sixth and seventh horizons. Judging from preliminary determinations, the taxonomic composition of ammonite assemblages clearly changes at this level. Foraminifers and conodonts are often encountered in deposits in addition to ammonoids. The uniform lithology, continuous sedimentation, and abundance of fossils set the section on Desli Caira Hill apart from other Olenekian–Anisian boundary sections in the Tethyan realm. It can be rightly considered as the most adequate candidate for the GSSP of the Anisian lower boundary. Nevertheless, ammonoids found in the section should be studied and described before its official approval.

The next step in the same discussion on the Middle Triassic stage boundaries at the international meeting held in Veszprém (Hungary), September 6–8, 2002 (*STS/IGCP 467 Field Meeting ...*, 2002). It was organized by the Triassic Subcommission of the Hungarian Stratigraphic Commission, the Hungarian Academy of Sciences, and by the Hungarian Geological Society on the STS's initiative within the framework of the IGCP Project 467 "The Triassic and Trans-Panthalassic Correlation." Besides nine Hungarian specialists, more than thirty scientists from Austria, Bulgaria, Great Britain, Spain, Italy, Canada, China, Poland, Russia, Slovakia, USA, Switzerland, Montenegro, Czechia, and Yugoslavia took part in that meeting.

The meeting began with one-day trip to sections of Anisian–Ladinian and Ladinian–Carnian boundary deposits in the eastern part of the Balaton Upland. The Felsöörs section was most interesting one among others suggested by Hungarian geologists the GSSP of the Ladinian lower boundary (Vörös *et al.*, 1996). Exposures on the Forräshegy Hill slopes are partially natural and partially stripped in man-made cuttings, protected against washout by special wooden sheds. Along the exposures declared as a natural geological monument, there is a sight road made up of crushed stone for tourists and school children. The Anisian–Ladinian boundary in this section is defined by Hungarian stratigraphers between the Felsöörs and Väszoly formations. The Felsöörs Formation is crowned by gray laminated limestones (less than 10 m thick) of the upper Anisian *trinodosus* Zone. The Väszoly Formation consists mainly of greenish tuffs and tuffites (20 m) intercalated with rare interlayers and lenses of yellowish limestone that yields ammonoids of the *reitzi* Zone. The zone is divided into the *Kellnerites felsoeoersensis, Hyparpadites liepoldti, Reitziites reitzi*, and *Aplococeras avisianum* subzones. Red siliceous limestones of the Buchenstein Formation occur above them. Boundary beds with *Ticinites* cf. *crassus* (Hauer), *Stoppaniceras* cf. *variabile* Rieber, and *Chieseiceras* sp. between the Väszoly and Buchenstein Formations are assigned to the *secedensis* Zone. A fragment of shell of *Eoprotrachyceras* sp. found above is indicative of the Ladinian *curionii* Zone. The Anisian–Ladinian boundary suggested by Hungarian stratigraphers in the Felsöörs section coincides with the lithological boundary, and this is inconsistent with international requirements to the quality of global boundary stratotypes.

Problems of the Middle Triassic stage boundaries were discussed during the next two days. Twenty oral and sixteen poster presentations were devoted to these problems. Sessions proceeded in a comfortable hall of the Regional Center of the Hungarian Academy of Sciences, which is housed in one of the old castles of Veszprém.

The first (morning) session of September 7 was devoted to the Olenekian–Anisian boundary. **M.J. Orchard**, the present chairman of the STS, was first to report on potential index forms of North American conodonts, which could be used for establishing the Middle Triassic stage boundaries and for correlation of sections in the Old and New World. Species *Chiosella timorensis* (Nogami) found along with ammonoids *Japonites welteri* Bucher in Anisian basal beds of Nevada and is of wide geographic range can be considered as the acceptable index fossil of the Olenekian–Anisian boundary. Species *Neogondolella* ex gr. *constricta* (Mosher et Clark), *Paragondolella* ex gr. *excelsa* (Mosher), and the genus *Budurovignathus* appear at the Anisian–Ladinian boundary defined between the ammonite *occidentalis–subasperum* zones in the USA and the *chischa–matutinum* zones in Canada. The Ladinian–Carnian boundary is defined between the ammonite *sutherlandi–obesum* zones in British Columbia and below the *desatoyense* Zone in Nevada. First representatives of the *Metapolygnathus* and *Mosherella* genera, and some morphotypes of the *Neogondolella* genus appear at the level of the *sutherlandi* Zone. *D. Lehrmann* and his colleagues reported on the results of comprehensive study of the Olenekian–Anisian boundary in the continuous Guandao section of marine deposits in the Nanpanjiang Basin of southern China. Species *Neospathodus homeri* (Bender) and *N. symmetricus* Orchard disappear at this boundary, while *Neospathodus gondolelloides* (Bender), *Chiosella timorensis* (Nogami), and *Neogondolella regalis* Mosher appear at this level separating rocks with reversed and normal geomagnetic polarities. The radiometric dates obtained for volcanic tuff beds in the studied section revealed ages of the Permian–Triassic and Olenekian–Anisian boundaries, which correspond to 251 and 247 Ma, respectively. Judging from these dates, the Early Triassic lasted no more than 4 Ma, and regeneration of marine biota after the Permian extinction advanced much faster than it was believed earlier. Chemostratigraphic studies carried out in different countries by *V. Atudorei* and his colleagues showed considerable changes in the carbon, sulfur, and stron-

Variants of Anisian-Ladinian boundaries and their interregional correlation

Southern Alps				Nevada			Canada	Siberia			
Lower Ladinian	Fassan	*Eoprotrachyceras*	*Xenoprotrachyceras recubariense*	Lower Ladinian	*Eoprotrachyceras subasperum*		*Eoprotrachyceras matutinum*				
			Eoprotrachyceras curionii								
– ? –	– ? –	*Nevadites*	*Chieseiceras chiesense*	Upper Anisian	*Frechites occidentalis*		*Frechites chischa*				
			Serpianites serpianensis								
			Ticinites crassus								
– ? –	– ? –	*Hungarites*	*Aplococeras avisianum*		*Parafrechites meeki*	*Parafrechites dunni*		*Frechites nevadanus*	*Parafrechites sublaqueatus*		
						Parafrechites meeki			*Frechites nevadanus*		
			Reitziites reitzi			*Frechites nevadanus*			*Parapopanoceras dzeginense*		
					Gymnotoceras rotelliforme	*Gymnotoceras blakei*		*Gymnotoceras rotelliforme*	*Gymnotoceras olenekense*		
– ? –	– ? –					*Paraceratites cricki*					
Upper Anisian	Illir	*Paraceratites*	*Paraceratites trinodosus*			*Paraceratites vogdesi*	*Eogymnotoceras deleeni*		*Parapopanoceras asseretoi*		
						Paraceratites clarkei					
			Schreyerites abichi			*Paraceratites burckhardti*					

tium isotopic ratios across the Olenekian–Anisian boundary. These geochemical indicators can be used for a high-resolution stratigraphic correlation. **E. Gradinaru and D. Ivanova** presented preliminary results on foraminiferal assemblages from Olenekian and Anisian boundary deposits in the mentioned Desli Caira section (Romania). **E. Gradinaru, M. Orchard, A. Nicora, E. Mirauta, and V. Atudorei** reported on conodont succession in the same deposits. **V.R. Lozovskii, D.A. Kukhtinov, and O.P. Yaroshenko** devoted their report to paleontologic and biostratigraphic characteristics of Lower–Middle Triassic continental and freshwater sequences of Eastern Europe. Their special attention was paid to positions of the Olenekian–Anisian and Anisian–Ladinian boundaries in the sequences.

The second (evening) session was devoted to the Anisian–Ladinian boundary, the most debatable one in the Triassic succession, as I think. Despite the general agreement that the boundary should be defined somewhere between the Alpine zones *trinodosus* (*Paraceratites* Genus-zone) and *curionii* (*Eoprotrachyceras* Genus-zone), the specialists were of controversial opinions on its concrete position within this stratigraphic range, placing it between the *Paraceratites* and *Hungarites*, *Hungarites* and *Nevadites*, or *Nevadites* and *Eoprotrachyceras* genus-zones (table).

The *Paraceratites–Hungarites* (or *trinodosus–reitzi*) boundary has a historical priority. E. Mojsisovics (1882) placed the lower boundary of his Norian Stage, renamed later into the Ladinian, at this level 120 years ago (Bittner, 1893). Until recently, this boundary with the stratotype in the Felsöörs section was unanimously recognized by Hungarian stratigraphers (Vörös *et al.*, 1996).

The base of the *Nevadites* Genus-zone is the second candidate for the Anisian–Ladinian boundary. This zone was distinguished 20 years ago by the Austrian geologist L. Krystyn (1983) in the Hallstatt Limestone of the Epidavros section (Greece) as the basal one in the Ladinian Stage. Following his opinion on stratigraphic position of the *Nevadites* Zone, Italian stratigraphers Mietto and Manfrin (1995) subdivided it in the Southern Alps into the *Ticinites crassus* (the Pian dei Fiaconno type area, Dolomites, Italy), *Serpianites serpianensis* (the Monte San Giorgio type area, Ticino Canton, Switzerland), and *Chieseiceras chiesense* (the Val Gola type section, Trento, Italy) subzones.

The base of the *Eoprotrachyceras* Genus-zone is the third possible level for the Anisian–Ladinian boundary. This genus-zone was introduced by Mietto and Manfrin (1995) who subdivided it in the Val Gola section into the *Eoprotrachyceras curionii* and *Xenoprotrachyceras recubariense* subzones. In the 1960s, Canadian geologist E.T. Tozer (Tozer, 1967; Silberling and Tozer, 1968) proposed to place the Ladinian lower boundary at the base of *Eoprotrachyceras* Beds of North America, which corresponds to the lower boundary of the *curionii* Zone (or Subzone) in the Alps. At the begin-

ning of the 1980s, the International Stratigraphic Commission of the USSR accepted this alternative of the Ladinian lower boundary (Zhamoida *et al.*, 1982). Later on, Brack and Rieber (1986, 1993) placed the boundary at the same level in sections of Brescian Prealps and Guidicarie (Italy).

According to Krystyn (1978), the appearance level of high-rank taxa (superfamilies or families) is an important indication the Triassic stage boundaries. Superfamily Trachyceratacеae is such a taxon of the Middle Triassic. The genus *Nevadites* is a stumbling block in this approach, because its status is debatable. Some specialists include it into the superfamily Ceratitaceae (Tozer, 1981; Shevyrev, 1986), whereas others attribute it to the superfamily Trachyceratacеae (Mietto and Manfrin, 1995). In addition, the first group of specialists considers the genus *Eoprotrachyceras* as the earliest representative of Trachyceratacеae, whereas their opponents believe the genus *Nevadites* was first in that superfamily.

Unfortunately, these disagreements have not been overcome at the Veszprém Meeting. Having found out that positioning of the Anisian–Ladinian boundary at the base of the *reitzi* s. l. Zone has no support, **A. Vörös** proposed to rise it in the Felsöörs section to the level of the synonymous subzone, were the appearance of zonal index species is recorded. In other words, he suggested to reconcile it with the base of the *Hungarites* Genus-zone. In his opinion, alternatives to place the Ladinian lower boundary at the base of the *avisianum* Subzone or at the base of the *secedensis* Subzone (*Nevadites* Genus-zone) are also plausible. As Vörös believes, the boundary at the base of the *curionii* Zone (*Eoprotrachyceras* Genus-zone) is the least acceptable, since its index species is rather rare. **H. Kozur** holds the same viewpoint. He also considers the *reitzi* s. str. Subzone as most adequate for the base of the Ladinian, because most significant changes in taxonomic composition of conodonts and radiolarians took place exactly at this level. Rather appreciable micropaleontological changes are also established at the lower boundary of the *avisianum* Zone. In distinction, **S. Kovács** came to the conclusion that the idea to place the boundary at the base of the *reitzi* s. str. Subzone (*Hungarites* Genus-zone) cannot be supported by any substantial event in the evolution of conodonts. In his opinion, it is more reasonable to place the boundary at the level of the *avisianum* Zone, where representatives of *Gondolella alpina* Kozur et Mosher and *G. trammeri* Kozur species appear. Kovács believes that stratigraphic ranges of conodonts, like of other fossils, are under strong facies control, and substantiating stage boundaries one should use a combined approach and consider different fossil groups with a considerable correlation potential. Tuff beds in the Felsöörs section allowed **J. Palfy, R. Parrish, and A. Vörös** to determine the lower age limits of the *reitzi* s. l. Zone (241.1 ± 0.5 Ma) and the *reitzi* s. str. Subzone (240.5 ± 0.5 Ma). Leaning upon the structure of lobe lines, **H. Rieber and P. Brack** showed that *Nevadites* and *Eoprotrachyceras* have no phylogenetic relations and assigned the genus *Nevadites* to Ceratitidae (superfamily Ceratitaceae). The genus *Eoprotrachyceras*, the first representative of superfamily Trachyceratacеae, is still of unclear origin.

Speaking at the session, I upheld the alternative of the Anisian–Ladinian boundary between the *Nevadites* and *Eoprotrachyceras* genus-zones (Shevyrev, 2002) based on the following arguments. First, the earliest true Trachyceratacеae appear at this level. Second, this boundary is of the greatest correlation potential being easily traceable everywhere. On the other hand, it is impossible to trace in some cases the other suggested levels of the Anisian–Ladinian boundary. For instance, they cannot be reliably established in sections of Canada (see table). **P. Brack, R. Mandil, G. Muttoni, H. Rieber, and W. Schatz** presented an overview and interpretation of results obtained by litho-, bio-, magneto-, and chronostratigraphic methods during the last decade of Anisian–Ladinian boundary investigation in the Southern Alps. The established marker beds and representative faunal horizons in the studied stratigraphic interval were used to reveal mutual correlation between numerous sections of the Southern Alps. The validity of the proposed litho- and biostratigraphic correlation is confirmed by datums of magnetic reversals. The correlation of strata between the upper part of the *reitzi* Zone and the *curionii* Zone is especially reliable. It clearly demonstrates a considerable difference in sedimentation rates on carbonate platforms and in abyssal basins. For instance, a 600-m-thick sequence of the Latemar Platform corresponds to a thin 25-m-thick succession of lower Buchenstein Beds in the Dolomites. According to radiometric data obtained by stratigraphers mentioned above, deposition of these beds lasted about 4 Ma, and the lower boundary of the *curionii* Zone taken for the base of the Ladinian is 240.7 Ma. They also revealed the most complete and continuous succession of macrofossils in the Badolino section (Brescian fore-Alps), which is proposed as a candidate for the GSSP of the Anisian–Ladinian boundary. According to data of **A. Nicora and P. Brack,** a complete succession of conodont assemblages is observable in this section. **N. Pretto, P. Mietto, and S. Manfrin** reported on stratigraphy of a thick carbonate sequence of the Latemar Platform (Dolomites). They distinguished 18 ammonite horizons corresponding to the *avisianum, crassus*, and *serpianensis* zones. The researchers defined the lower boundary of the Ladinian at the base of the *crassus* Zone (*Nevadites* Genus-zone). They also noted a considerable distinction between sedimentation rates on carbonate platforms and in basins, which should be taken into account by selection of the GSSP for the Anisian-Ladinian boundary. After analysis of palynological assemblages from one section in Dolomites, **P. Hochuli and G. Roghi** concluded that the greatest taxonomic changes are recorded between the *reitzi* Zone (*Hungarites* Genus-

zone) and the *curionii* Zone (*Eoprotrachyceras* Genus-zone).

During the third (conclusive) session, held on September 8, the Ladinian–Carnian boundary was under debates. **M. Gaetani** reported on the relevant general problems. He recalled that up to the mid-1990s it was the most satisfactory stage boundary in the Triassic. It was unanimously defined at the appearance level of *Trachyceras aon* (Münster) in the Alps and of *T. desatoyense* Johnston in North America. It was a general opinion that the genus *Trachyceras* also appeared at this level. However, Mietto and Manfrin (1995) showed that rare representatives of that genus, for instance, *T. bipunctatum* (Münster) and *T. muensteri* (Wissmann), appear in Dolomites earlier than *T. aon* and suggested to define the Ladinian–Carnian boundary at the appearance level of the genus *Daxatina*, i.e. below the first occurrence of *T. aon*. They included the *Daxatina* cf. *canadensis* Subzone, as a basal subzone of the Carnian, into the *Trachyceras* Genus-zone. Soon this point of view was supported by other Italian stratigraphers (Broglio Loriga *et al.*, 1999), who also established the boundary position at the base of *Daxatina* Beds and proposed the section of the San Cassiano Formation in Prati-di-Stuores (Dolomites, west of Cortina d'Ampezzo) for its GSSP. Contrary to the genus *Trachyceras*, the geographic range of which is confined to lower and middle latitudes, *Daxatina* is a cosmopolitan genus and, hence, is more suitable for global correlation. **M. Balini, L. Krystyn, A. Nicora, and V. Torti** reported on the results of their study of Ladinian–Carnian boundary deposits in the Spiti Valley (Central Himalayas). They distinguished in these deposits five successive ammonoid assemblages named after index genera *Meginoceras, Maclearnoceras, Frankites, Daxatina*, and *Trachyceras*. Stratigraphic ranges of genera *Frankites* and *Daxatina* are slightly overlapping here. They also found it difficult to correlate the appearance levels of genus *Trachyceras* in the five studied sections. The most important event in evolution of pelagic bivalves is the appearance of genus *Halobia* in the middle part of stratigraphic range of the genus *Frankites*. First conodont species *Metapolygnathus polygnathiformis* (Budurov et Stefanov) appear at the same level. Some participants of the International Conference in Lausanne (1991) suggested the Ladinian–Carnian boundary also at this level in the Balaton Uplift sections. The reported data showed that the most important events in evolution of bivalves and conodonts do not coincide with indicative moments in development history of ammonoids. The studied Himalayan sections can be regarded as adding some interesting materials to the Prati-di-Stores Section in the Alps. **G. Warrington** gave characterization of Triassic deposits, which are exposed in the Dorset and East Devon cliffs. This locality is nominated as a World Heritage Site by UNESCO in 2001. Exposed here is a thick sequence comprising deposits from barren continental Aylesbeare Mudstone of the Upper Permian?–Lower Triassic to marine shale of the Penarth Group with Rhaetian foraminifers, bivalves, gastropods, corals, conodonts, crustaceans, and fish remains. The sequence is overlaid by the Lias Group, in which *Psiloceras planorbis* Sow., the index species of the Jurassic basal zone, appears 2.5 m above its base. The pre-*planorbis* beds of the group, which yield foraminifers, brachiopods, bivalves, ostracodes, and echinoids, are regarded as the uppermost Triassic deposits. Concluding at the session was the report by **M.A. Shishkin and V.G. Ochev** on Middle Triassic tetrapod faunas of Eastern Europe and on problems of their dating.

The selection of Global Stratotype Sections and Points for Triassic stage boundaries will be continued. For this purpose, the staff of international working groups on Carnian–Norian and Norian–Rhaetian boundaries has been changed. These boundaries are planed to be discussed on May 26–28, 2003, at the session in Vancouver (Canada) devoted to biotic extinctions, faunal changes, and natural boundaries in the Upper Triassic, the classic sections of which will be visited in the vicinity of Williston Lake, northeast of British Columbia (Orchard *et al.*, 2001). The discussion of the same boundaries is planned to be continued at the workshop in Venice (Italy), August, 2004, after observation of some Upper Triassic sections in the Alps, which are surmised to be global boundary stratotypes.

Reviewers E.Yu. Baraboshkin and V.V. Mitta

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